Bulletin

of

The Geological Society of America

VOLUME 66 January 1955 NUMBER 1

CONTENTS

	Pages
Regimen and flow of ice in equilibrium glaciers. By Lawrence E. Nielsen	1–8
Trondhjemite batholith near Bidwell Bar, California. By Robert R. Compton	9–44
Glendora volcanic rocks, Los Angeles Basin, California. By John S. Shelton	45–90
Peridotites of Manpur, Singhbhum District, Bihar, India, and origin of associated asbestos deposits. By S. C. Chatterjee	91–104
Geology of the Hanover quadrangle, New Hampshire-Vermont. By J. B. Lyons	105–146
Short Notes	
Thickness of bedding and parting in sedimentary rocks. By Henry H. Gray	147–148
Preliminary report on the system SiO_2-H_2O. By O. F. Tuttle and J. L. England	149–152
Selected bibliography of standard stratigraphic sections of North America. By Fred S. Honkala	153–154
Paleoecological significance of the strontium-calcium ratio in fossils and sediments. By K. Turekian	155–158

Subscription $15.00 per year.

Printing Office: Mt. Royal & Guilford Aves., Baltimore 2, Md.

Communications for publications should be addressed to The Geological Society of America, Dr. H. R. Aldrich, Secretary, 419 West 117 Street, New York, N. Y.

Please give notice of change of address 4 weeks in advance. Claims for non-receipt of the Bulletin of a given month must be sent to the Secretary of the Society before the end of the second succeeding month.

Entered as second-class matter at the Post Office at Baltimore, Md.,
under the Act of Congress of July 16, 1894.
Accepted for mailing at special rate of postage provided for in Section 1103
Act of October 3, 1917, authorized on July 8, 1918.

PAPERS IN PRESS FOR FORTHCOMING ISSUES

GEOLOGY OF CATOCTIN MOUNTAIN, MARYLAND AND VIRGINIA. By John C. Whitaker

A geologic map, in color, accompanies this paper. Separate copies of the map, unfolded and in tubes, may be purchased. Price 25 cents apiece. Remittance should accompany order.

SAMPLING IN BIOGEOCHEMISTRY. By H. V. Warren and Fortescue Delavault

SOURCE OF SHARON CONGLOMERATE OF NORTHEASTERN OHIO. By J. Osborn Fuller

TOPOGRAPHY OF THE CONTINENTAL SLOPE IN NORTHWEST GULF OF MEXICO. By Betty Lee Gealy

SEDIMENTARY FACIES AND INTERTONGUING IN THE UPPER CRETACEOUS, BOOK CLIFFS, UTAH-COLORADO By Robert G. Young

GEOLOGY OF THE LIMA REGION, SOUTHWESTERN MONTANA AND ADJACENT IDAHO. By Robert Scholten, K. A. Keenmon, and W. O. Kupsch

A geologic map, in color, accompanies this paper. Separate copies of the map, unfolded and in tubes, may be purchased. Price 25 cents apiece. Remittance should accompany order.

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 66, PP. 1-8, 2 FIGS. JANUARY 1955

REGIMEN AND FLOW OF ICE IN EQUILIBRIUM GLACIERS

BY LAWRENCE E. NIELSEN

ABSTRACT

An equilibrium glacier is one that has constant dimensions, since the average net accumulation per budget year above the annual firn limit exactly equals the amount of bare ice lost each year by ablation below the firn limit; the flow of ice maintains the glacier in exact balance over a period of years. Three basic equations that apply to an equilibrium glacier are:

$$Q_1 = \int_0^X H\,dX$$

$$Q_2 = Q^* - \int_A^X H'\,dX$$

$$\bar{v} = Q/S\rho$$

The first equation gives the mass of ice Q_1 moving through any cross section in the accumulation area in terms of the mass of snow H at any point X along the glacier at the end of the summer season. The second equation gives the mass of ice Q_2 moving through any cross section in the ablation region in terms of the mass of ice Q^* passing through the firn line and the ablation H' of bare ice per year. The third equation shows how the average velocity of flow $\bar{v}$ depends upon the mass of ice Q through a cross section of area S and ice of density ρ.

These equations must be solved by numerical approximations for an actual glacier, but they may be solved exactly for idealized glaciers of many shapes; this has been done for eight idealized examples. The results show that the velocity of flow is usually greatest near the annual firn limit and very small near the head and the terminus.

The theory predicts longitudinal expansion in the accumulation area and longitudinal contraction in the ablation area. Transverse surface profiles of equilibrium glaciers are generally higher in the center than at the edges in the ablation area, but in the accumulation area the transverse profiles are generally dish-shaped.

Large rates of ablation require large velocities of flow in order to maintain equilibrium. This concept is used to explain some of the phenomena often observed near ice-dammed lakes.

CONTENTS

TEXT

	Page
Introduction	1
List of symbols and definitions	2
Mathematics of equilibrium glaciers	3
Illustrative calculations on glaciers	4
Other implications of the theory	8
References cited	8

ILLUSTRATIONS

Figure	Page
1. Schematic diagram of an idealized glacier	3
2. Mass discharge and flow characteristics of idealized glaciers and rates of accumulation or ablation	7

INTRODUCTION

Although most glaciologists are familiar with mass transfer of ice in glaciers in terms of rates of accumulation and ablation and the dimensions of a glacier, no detailed and quantitative description has been published of mass balance and movement of ice in equilibrium glaciers. Incomplete descriptions have been given by Réné Koechlin (1944), Hans Hess (1904, p. 345), and G. S. Callendar (1950, p. 459). Reid (1896, p. 912) discussed many of the phenomena described in this paper qualita-

tively but not quantitatively.[1] Robert Sharp (1951, p. 726) has recently applied a similar concept of regimen to his experimental study of the Seward-Malaspina system.

A glacier is in a state of equilibrium when it is not advancing or retreating and not getting any thicker or thinner throughout its length. The excess accumulation above the firn line equals the amount of ice lost by ablation below the firn line, averaged over several years. This quantity of ice is transported from the region of accumulation to the region of ablation by glacial flow. The net accumulation of snow at the end of the ablation season is important, not the total snowfall which has little to do with the state of a glacier or the amount of ice that will be moved. Likewise, in the lower regions of a glacier the quantity of ice lost by ablation processes after removal of the current year's snowfall is important in studies of mass balance, as is the average position of the firn line or firn limit at the end of the ablation season; this line marks the point where the total ablation and accumulation are exactly balanced.

This work was carried out under a Post-Doctorate Research Fellowship at Harvard University, Cambridge, Mass., 1952–1953, and was given in part at a meeting of the American Geophysical Union in Washington, D. C. on May 3, 1954.

[1] The author wishes to thank Prof. Robert Sharp for bringing this paper to his attention after the completion of this work.

List of Symbols and Definitions

X Distance along glacier starting with the upper end at $X = 0$.
A Distance from the beginning of a glacier to its average firn limit. $X = A$ at the firn line. For a circular ice cap it is the distance from the center to the firn limit.
B Total length of a glacier. $X = B$ at the terminus. For a circular ice cap the distance from the center to the edge is B.
C Surface width of a glacier at any given point along the glacier.
C_0 Surface width at firn line.
C_1 Surface width of a glacier at its head (near bergschrund).
C_2 Surface width at terminus.
D Thickness of ice at any given point.
D_0 Maximum thickness of ice along any given cross section of a glacier.
Q Mass of ice passing through a given cross section in unit time, i. e., discharge per year.
Q_1 Mass of ice passing through a cross section in the accumulation region in a unit of time such as a year.
Q_2 Mass of ice passing through a cross section per year in the ablation region.
Q^* Mass of ice flowing past the position of the average firn limit per year.
V Volume of ice moving through a given cross section in a year.
ρ Density of ice in the glacier.
S Cross sectional area at any point.
$\bar{v}$ Average velocity of flow of ice at any point as defined by $\bar{v} = V/S$. (A bar over a symbol indicates an average.)
v_0 Velocity of flow of ice on the surface.
v_b Velocity of movement at the bed of the glacier, i. e., slippage along the bottom.
H Mass of excess accumulation during an "average" year for a strip of unit width across the glacier above the firn limit.
h Mass of the average yearly net accumulation per unit of area above the firn limit. Subscript zeros on H or h designate maximum values usually found at the head of the glacier.
H' Net mass lost by ablation from a strip of unit width across the glacier below the firn limit during an "average" year.
h' Net mass lost by ablation per unit of area below the firn limit during an "average" year. Subscript zeros on H' or h' designate maximum values of ablation usually found at the terminus.
K A constant which is characteristic of the shape of a cross section. Normal values of K lie between 0.6 and 1.0.
R Radial distance from the center of a circular ice cap to any given point.
η A quantity describing the flow characteristics of ice. In special cases η is the viscosity.
n A constant describing the flow characteristics of ice. $n \simeq 3$ for glacier ice.
τ Shear stress in ice.
θ Angle of surface slope.
g Acceleration of gravity.

Mathematics of Equilibrium Glaciers

An idealized glacier is illustrated in Figure 1. Quantity X indicates the distance along the length of the surface. The glacier originates near the bergschrund where $X = 0$. The late summer firn line occurs at $X = A$, while the

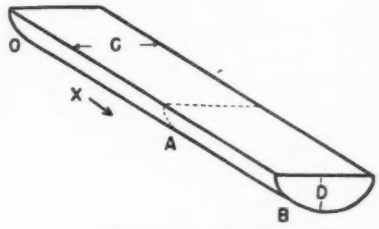


Figure 1.—Schematic Diagram of an Idealized Glacier

terminus is at $X = B$. The region of accumulation is between 0 and A; the region of ablation is between A and B. The fundamental assumption for equilibrium is that the mass of excess accumulation should equal the mass lost by ablation. Thus, the quantity of ice moving through a cross sectional area will be a maximum at the firn line and will be zero at both ends of the glacier (Reid, 1896, p. 912). However, the velocity of glacial flow at any given cross section is determined by the volume of ice passing through that cross section, so the velocity is not necessarily maximum at the firn line. The volume is related to the mass by the density of the ice.

Let Q be the mass of ice passing through a given cross section in a unit time, *i.e.*, Q is the mass discharge per year. Since Q depends upon at what part of the glacier it is measured, Q is a function of X. The quantity V is the volume of ice going through a given cross section in a unit time. Thus Q and V are related by the equation

$$V = Q/\rho \qquad (1)$$

where ρ is the density of the ice. (This density is the average value in the interior of the glacier, not at the surface.) Volume is also equal to the cross sectional area S at any point multiplied by the average velocity of flow $\bar{v}$ at that point:

$$V = S\bar{v} \qquad (2)$$

Combining equations (1) and (2) gives

$$\bar{v} = V/S = \frac{Q}{S\rho} \qquad (3)$$

Average velocity $\bar{v}$ is less than the average surface velocity $\bar{v}_0$.

Let H be the mass of excess accumulation for an "average" year for a strip of unit width across the glacier above the firn limit. This mass is equal to the product of the excess accumulation h per unit of area and the width of the glacier C at its surface. By excess accumulation is meant the total snow, frozen rain, and refrozen melt water minus the total ablation during a typical budget year.

Both h and C can depend upon X:

$$H = hC \qquad (0 \leq X \leq A) \quad (4)$$

Similarly H' is the mass lost by ablation from a strip of unit width across the glacier below the firn limit. It equals the loss h' per unit of area for a given value of X multiplied by the width of the glacier at that point.

$$H' = h'C \qquad (A \leq X \leq B) \quad (5)$$

This ice is the yearly quantity lost after removal of the snow cover and other forms of accumulation of the current year. This ablation is not necessarily the amount found during a given year but should be the average of several years.

The mass of ice passing through any given cross section in unit time in the accumulation region is

$$Q_1 = \int_0^X H\,dX \qquad (0 \leq X \leq A) \quad (6)$$

For equilibrium the amount of ice passing through a cross section per year must equal the total r t accumulation above this cross section per year as averaged over several years. This basic equation applies any place between the bergschrund and the firn line (between $X = 0$ and $X = A$). In the ablation region the mass of ice passing through a given cross section is:

$$Q_2 = Q^* - \int_A^X H'\,dX \qquad (A \leq X \leq B) \quad (7)$$

Here Q^* is the quantity of ice moving through the firn line per year; more ice passes through

this point than through any other point along the glacier.
The integral

$$\int_A^X H' \, dX$$

represents the total amount of ice lost because of ablation processes between the points A and X. Equations (6) and (7) are exact for an equilibrium glacier, and they point out the significance of the position of the firn line in determining the size and behavior of a glacier.

Illustrative Calculations on Glaciers

In actual cases, the integrals in equations (6) and (7) will have to be solved by numerical and approximation methods since most glaciers have complicated shapes and complex distribution of snowfall and rates of ablation. However, these equations may be solved exactly for certain idealized shapes of glaciers. This will be calculated in detail for a valley glacier whose width, C, is constant throughout its entire length. The yearly excess accumulation will be assumed to be greatest at the head of the glacier ($X = 0$) diminishing linearly to zero at the firn line. The yearly rate of excess ablation is assumed to be zero at the firn line, increasing linearly to the terminus of the glacier ($X = B$). These conditions could be realistic for some glaciers. Therefore:

$$H = H_0(1 - X/A)$$

and

$$H' = H_0'\left(\frac{X - A}{B - A}\right) \tag{8}$$

$$H_0 = hC$$

and

$$H_0' = h'C.$$

The quantity H_0 is the yearly excess amount of snow for a strip of unit width across the head of the glacier. H_0' is the yearly loss of ice by ablation for a strip of unit width across the terminus of the glacier. Thus, equations (6) and (7) can be solved to give:

$$Q_1 = \int_0^X H_0(1 - X/A)\, dX = H_0X\left(1 - \frac{X}{2A}\right) \tag{9}$$

in the accumulation area, and

$$Q_2 = Q^* - \int_A^X H_0'\left(\frac{X - A}{B - A}\right) dX = Q^* - \frac{H_0'}{2(B - A)}(X - A)^2 \tag{10}$$

in the ablation region.

Since

$$Q^* = \frac{H_0A}{2}, \tag{11}$$

$$Q_2 = \frac{H_0A}{2} - \frac{H_0'}{2(B - A)}(X - A)^2 \tag{12}$$

Also since

$$Q_2 = 0 \text{ at } X = B,$$

$$H_0' = H_0\left(\frac{A}{B - A}\right). \tag{13}$$

Case 2 of Figure 2 shows that the quantity of ice passing through a cross section changes along the glacier. Quantity Q is zero at both ends of the glacier and is a maximum at the firn line. As will be shown, the velocity of flow as measured on the surface will form a similar curve when plotted against X.

Equation (13) shows that the position of the firn line and the length of a glacier depend upon the rates of accumulation and ablation. For instance, a glacier with a high rate of accumulation (large H_0) and a small rate of ablation (small H_0') would have most of its length below the firn line. That is, B would be several times as large as A.

By the use of equation (3), it is possible to change the variable from the mass of ice moved to velocity of flow and the thickness of the ice. The density used in this equation is an average for any cross section. In the lower end of the glacier this density will approach the value for pure ice (0.9). In the accumulation region the density will be less, but still much greater than that of the surface snow since the average density for the mass of ice within the glacier is the important value. In general the density ρ will be between 0.8 and 0.90 and may be considered nearly constant throughout the length of the glacier. The value of the cross sectional area S also enters into equation (3); this is often unknown since the depth of the ice and the shape

of the cross section may not be known. However, the cross sectional area can be expressed approximately in terms of the width of the glacier at its surface, the maximum thickness of the ice D_0, and a constant K which depends only upon the shape of the cross section.

$$S = CD_0K \tag{14}$$

For semicircular and semielliptical cross sections $K = \pi/4$, while for parabolic cross sections $K = 2/3$. Therefore, in general, K has a value of approximately 0.7, but in the limit of a thin but wide glacier K approaches 1 as a limit. Equation (3) now becomes

$$\bar{v}D_0 = \frac{Q}{K\rho C} \tag{15}$$

while equations (9) and (10) become

$$\bar{v}D_0 = \frac{H_0X(1 - X/2A)}{K\rho C} \quad (0 \leq X \leq A) \tag{16}$$

$$\bar{v}D_0 = \frac{Q^*}{K\rho C} - \frac{H_0'(X - A)^2}{2K\rho C(B - A)} \quad (A \leq X \leq B). \tag{17}$$

Equations (16) and (17) show how the average velocity of flow, the thickness of the ice, the rates of ablation and accumulation, and the dimensions of a glacier are all related in a glacier of uniform width.

The usual practice in studying the flow of glaciers has been to put rows of movement stakes across a glacier. Something may also be learned by placing rows of stakes longitudinally along a glacier. In the case being used as an illustration, differentiation of equations (16) and (17) gives

$$\frac{d\bar{v}}{dX} = \frac{H_0}{K\rho CD_0}\left[(1 - X/A) - \frac{X(1 - X/2A)}{D_0}\frac{dD_0}{dX}\right] \tag{18}$$

in the accumulation area, and

$$\frac{d\bar{v}}{dX} = \frac{1}{K\rho CD_0}\left[\left(\frac{H_0'(X - A)^2}{2(B - A)D_0} - \frac{Q^*}{D_0}\right)\frac{dD_0}{dX} - \frac{H_0'(X - A)}{B - A}\right] \tag{19}$$

in the ablation area. These equations show that in the névé region, stakes placed longitudinally along the glacier will in time get farther and farther apart unless the ice thickens rapidly as the terminus is approached. Likewise, below the firn line the stakes will get closer and closer together with time unless the ice thins very rapidly as it moves down glacier. Such measurements made near the head and the terminus of a glacier are a theoretical way of determining the average rates of accumulation and ablation.

Equations (6) and (7) can be solved for glaciers (or parts of glaciers) of other shapes and rates of accumulation or ablation, such as the following (Fig. 2):

CASE 1.—The width is constant. The rate of accumulation H is constant above the firn line, and the rate of ablation H' is constant below this line. Then:

$$Q_1 = HX$$

and

$$Q_2 = HA - H'(X - A) = HA\left(1 - \frac{X - A}{B - A}\right).$$

In all cases the Q_1 equation applies for values of X between zero and A while the Q_2 equation applies for values of X between A and B.

CASE 2.—The width is constant. Rate of accumulation is a maximum H_0 at the head of the glacier and decreases to zero at the firn line. The rate of ablation is zero at the firn line and increases linearly to a maximum H_0' at the terminus. This case was discussed in detail above.

CASE 3.—The névé area is in the shape of a truncated cone with the small end down glacier. The rate of accumulation h per unit of area is constant. Then

$$C = C_1 - \left(\frac{C_1 - C_0}{A}\right)X$$

where C_1 is the width of the upper end of the glacier and C_0 is the width at the firn line. Below the firn line the glacier can be any shape. The accumulation area and the ablation area can be treated independently of one another:

$$Q_1 = h\left[C_1X - \left(\frac{C_1 - C_0}{2A}\right)X^2\right].$$

CASE 4.—The accumulation area is a truncated cone as in Case 3, but the amount of accumula-

tion ranges from a maximum at the head of the glacier to zero at the firn line:

$$C = C_1 - \left(\frac{C_1 - C_0}{A}\right) X$$

and

$$h = h_0(1 - X/A).$$

For this case

$$Q_1 = h_0 X\left[C_1 - \left(\frac{2C_1 - C_0}{2A}\right)X + \left(\frac{C_1 - C_0}{3A^2}\right)X^2\right].$$

CASE 5.—The ablation area is a truncated cone with the small end at the firn line, and h', the loss per unit area, is constant. These conditions could be typical of certain piedmont glaciers:

$$C = C_0 + \left(\frac{C_2 - C_0}{B - A}\right)(X - A)$$

where C_0 is the width at the firn line, and C_2 is the width at the terminus.

$$Q_2 = Q^* - h'$$

$$\left[C_0(X - A) + \frac{C_2 - C_0}{2}(B - A)\left(\frac{X - A}{B - A}\right)^2\right].$$

CASE 6.—Ablation area is shaped like a truncated cone as in Case 5, but the ablation per unit of area increases down glacier to a maximum of h_0' at the terminus.

$$h' = h_0'\left(\frac{X - A}{B - A}\right)$$

$$Q_2 = Q^* - h_0'(B - A)$$

$$\left(\frac{X - A}{B - A}\right)^2\left[\frac{C_0}{2} + \frac{C_2 - C_0}{3}\left(\frac{X - A}{B - A}\right)\right].$$

CASE 7.—The ablation area is cone-shaped with the point down glacier, and the ablation rate per unit area increases linearly down glacier. These conditions could be typical of some valley glaciers that end in pointed tongues.

$$C = C_0\left(1 - \frac{X - A}{B - A}\right), \quad h' = h_0'\left(\frac{X - A}{B - A}\right)$$

For this case

$$Q_2 = Q^* - h_0' C_0(B - A)$$

$$\left[\frac{1}{2}\left(\frac{X - A}{B - A}\right)^2 - \frac{1}{3}\left(\frac{X - A}{B - A}\right)^3\right].$$

CASE 8.—For a circular ice cap the rate of accumulation per unit area is assumed to be a maximum at the center and decreases linearly to zero at the firn line. The rate of ablation per unit area increases from zero at the firn line to a maximum at the edge of the ice cap.

$$h = h_0\left(\frac{A - R}{A}\right), \quad h' = h_0'\left(\frac{R - A}{B - A}\right)$$

where R is the distance from the center, and A is the distance from the center to the firn line.

For the inner portion of the ice cap

$$Q_1 = \frac{2\pi h_0}{A}\left(\frac{AR^2}{2} - \frac{R^3}{3}\right).$$

For the outer ablation area

$$Q_2 = \pi/3\left[h_0 A^2 - \frac{h_0'}{B - A}(2R^3 - 3AR^2 + A^3)\right].$$

These cases and others can be combined in many ways to describe glaciers of different shapes and characteristics, both above and below the firn line. Only in exceptional cases would an actual glacier exactly fulfill the conditions for shape, accumulation, and ablation, but these cases do illustrate some of the possible variations.

In Figure 2, the second column gives an aerial view, and the third column gives the quantity of ice Q through any cross section plotted as a function of distance along the glacier for the 8 cases described above. The last column gives the product of the average velocity $\bar{v}$ and thickness of the ice D_0 as a function of distance along the glacier. It is impossible to separate further the variables $\bar{v}$ and D_0 unless one of them is known or is calculated independently from other information. The quantity K may also be unknown, but it is a very slowly changing function and can be approximated. Corresponding curves for Q and $\bar{v}D_0$ generally have similar shapes; however, they may differ considerably when the dimensions of the glacier are changing rapidly, and the maximum velocity of flow may not occur exactly at the firn line where Q is always a maximum. The position of the firn line is indicated by a vertical dashed line in Figure 2.

In the discussion so far, an average velocity

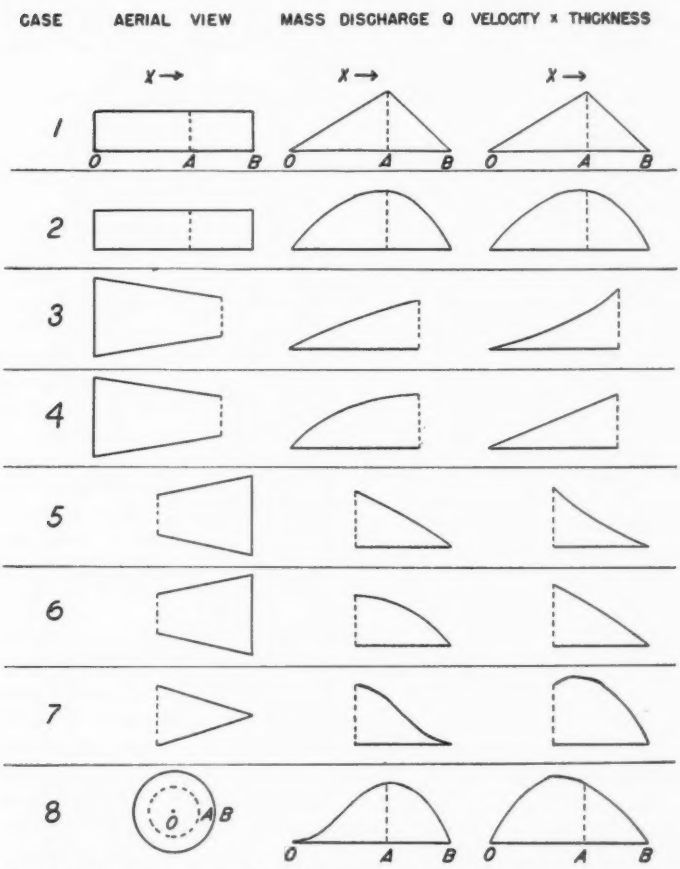


FIGURE 2.—MASS DISCHARGE AND FLOW CHARACTERISTICS OF IDEALIZED GLACIERS OF VARIOUS SHAPES AND RATES OF ACCUMULATION OR ABLATION

The position of the firn line is indicated by a dashed line. Cases 3 and 4 show only the accumulation area of the glacier, while cases 5, 6, and 7 show only the ablation area.

of flow, defined as $\bar{v} = V/S$, has been used. In terms of such variables as surface velocity, this average is really a double average which takes into account the variation in surface velocity across the glacier and the variation in velocity through the thickness of the ice. This average velocity is in reality much more difficult to measure than the surface velocity. On the basis of the theory of glacier mechanics worked out by Nye (1952, p. 82), the average bulk velocity so important to the theory can be estimated from the surface velocity, at least in the case of a glacier whose width is many times its thickness.

After Nye (1952, p. 83),

$$\frac{dv}{dD} = \left(\frac{\tau}{\eta}\right)^n \tag{20}$$

where

$$\tau = \rho g D \sin \theta \tag{21}$$

Equation (20) shows that the velocity changes with depth, while equation (21) shows that the shearing stress τ depends upon the surface slope angle θ. The quantities n and η are dependent upon the properties of ice and are supposedly constant at a given temperature;

η is analogous to the viscosity of ice; g is the acceleration of gravity.

Equation (20) can be integrated to give (see Nye's equation 3):

$$v_0 - v_b = \left(\frac{\rho g}{\eta}\right)^n \frac{D_b^{n+1}}{n+1} \sin^n \theta \qquad (22)$$

where v_0 is the surface velocity, v_b is the velocity at the bottom of the glacier (slippage), and D_b is the thickness of the ice. Substitution of equation (22) (Nye's equation 3) into Nye's equation 2 gives

$$v = \left(\frac{\rho g}{\eta}\right)^n \left[\frac{D_b^{n+1} - D^{n+1}}{n+1}\right] \sin^n \theta + v_b \qquad (23)$$

Integration of equation (23) with respect to D and rearrangement gives Nye's equation 4 for the volume. Division of the volume equation by D_b gives the following equation as a special case for the desired average velocity of flow $\bar{v}$,

$$\bar{v} - v_b = \left(\frac{\rho g}{\eta}\right)^n \frac{D_b^{n+1}}{n+2} \sin^n \theta \qquad (24)$$

Combining equations (22) and (24) gives the relation between the average velocity and the surface velocity,

$$\frac{\bar{v} - v_b}{v_0 - v_b} = \frac{n+1}{n+2}. \qquad (25)$$

Since n is about 3 for ice (Glen, 1953, p. 721), the average velocity is generally 80–100 per cent of the surface velocity.

Other Implications of the Theory

The theory implies that in the ablation region of a glacier the transverse surface profile should be dome-shaped—that is, the ice should be higher at the center than at the edges. This is apparent, since the equilibrium condition holds for any portion of the glacier and is just as applicable across a glacier as up and down it. In a strip across the glacier in the ablation area as much ice must move in as is lost by ablation at any point. In general the velocity of flow and velocity gradient are greater in the center than near the edges. Therefore, even where the rate of ablation is no greater near the edges than at the center, the flow of ice is not straight down the glacier but trends toward the edges in order to maintain the balance. In order to have ice flow from the center to the edges, the center must be higher. Thus, in many glaciers medial moraines are found closer to the edge as the terminus is approached. In the accumulation area the opposite should be true; a transverse profile should be dish-shaped so that the ice moves not only down glacier but also toward the center.

Many ice-dammed lakes illustrate the same thing. The water of a lake acts as a powerful erosion medium. Buoyancy tends to break off icebergs; a water surface absorbs much more energy from sunlight than does ice or snow and this heat energy can melt large amounts of ice so the edge of the lake is a region of extremely high ablation. Abnormally large quantities of ice must flow toward the lake to compensate for this melting and for erosion brought about by the lake if equilibrium is to be maintained. Very often a glacier does not flow by an ice-dammed lake but a large portion of it turns and flows toward the lake even though this may require a drastic change in direction. This is another indication that, to maintain a steady state or equilibrium condition, the ice must flow in such a manner as to compensate for the ice lost by the various ablation processes.

Thus even though most glaciers are not in true equilibrium, the general theory does help to understand glacial movement and should be at least qualitatively applicable to glaciers that are not too far from an equilibrium state.

References Cited

Callendar, G. S., 1950, Relation between the height of the firn line and the dimensions of a glacier: Jour. Glaciology, v. 1, p. 459–461.

Glen, J. W., 1953, Rate of flow of polycrystalline ice: Nature, v. 172, p. 721–723.

Hess, H., 1904, Die Gletscher: Braunschweig, Germany, Friedrich Vieweg und Sohn, p. 345.

Koechlin, R., 1944, Les glaciers et leur mécanisme: Lausanne, France, F. Rouge and Co.

Nye, J. F., 1952, The mechanics of glacier flow: Jour. Glaciology, v. 2, p. 82–93.

Reid, H. F., 1896, The mechanics of glaciers: Jour. Geology, v. 4, p. 912–928.

Sharp, R. P., 1951, Accumulation and ablation on the Seward-Malaspina glacier system, Canada-Alaska: Geol. Soc. America Bull., v. 62, p. 726–744.

Monsanto Chemical Co., Plastics Division, Springfield 2, Mass.

Manuscript Received by the Secretary of the Society, April 2, 1954

124° 30′ 25′

EXPLANATION

TERTIARY ROCKS

Basaltic lava

LATE JURASSIC ROCKS

Aplite and fine-grained granitic rocks

Aplite and pegmatite dike swarms

Granophyre

(1) (2) (3) (4)

Granitic rocks, showing massive and flow layered types: (1) tonalite, (2) granodiorite, (3) trondhjemite, and (4) leucotrondhjemite
(12) = specimen no.

(1) (2)

Tonalite in migmatites: (1) chiefly injection type, (2) chiefly replacement type

Heterogeneous gabbroic and dioritic rocks

LATE PALEOZOIC (?) ROCKS

Metagranodiorite porphyry

Metagabbro and metadiorite

Metaserpentine

(1) (2)

Metavolcanic rocks of contact aureole: (1) fine-grained hornfels, and (2) coarse-grained hornfels and amphibolite

(1) (2)

Low grade metabasaltic rocks: (1) schistose, and (2) massive

Metarhyolite and metadacite

Metasedimentary rocks, chiefly non-calcareous

0 ½ 1 Mile

0 5000 Feet

TRUE NORTH 18°

Principal roads Principal streams

Culture and drainage abstracted from 1947 7½′ topographic maps of the U. S. Bureau of Reclamation

37° 35′

37° 30′

121° 30′ 25′

Migmatized metadacite

Hare Creek

Canyon

Bidwell Bar

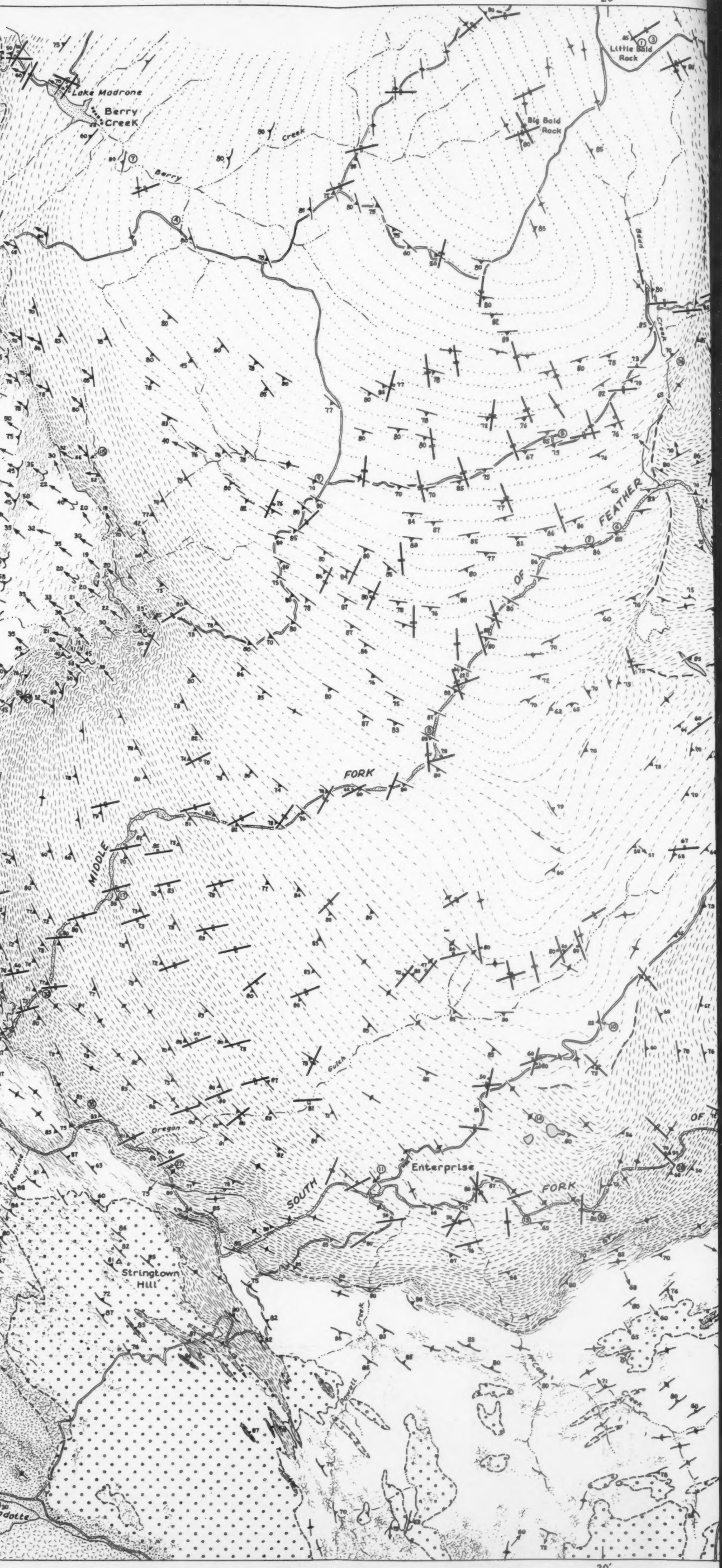
Bidwell Junction

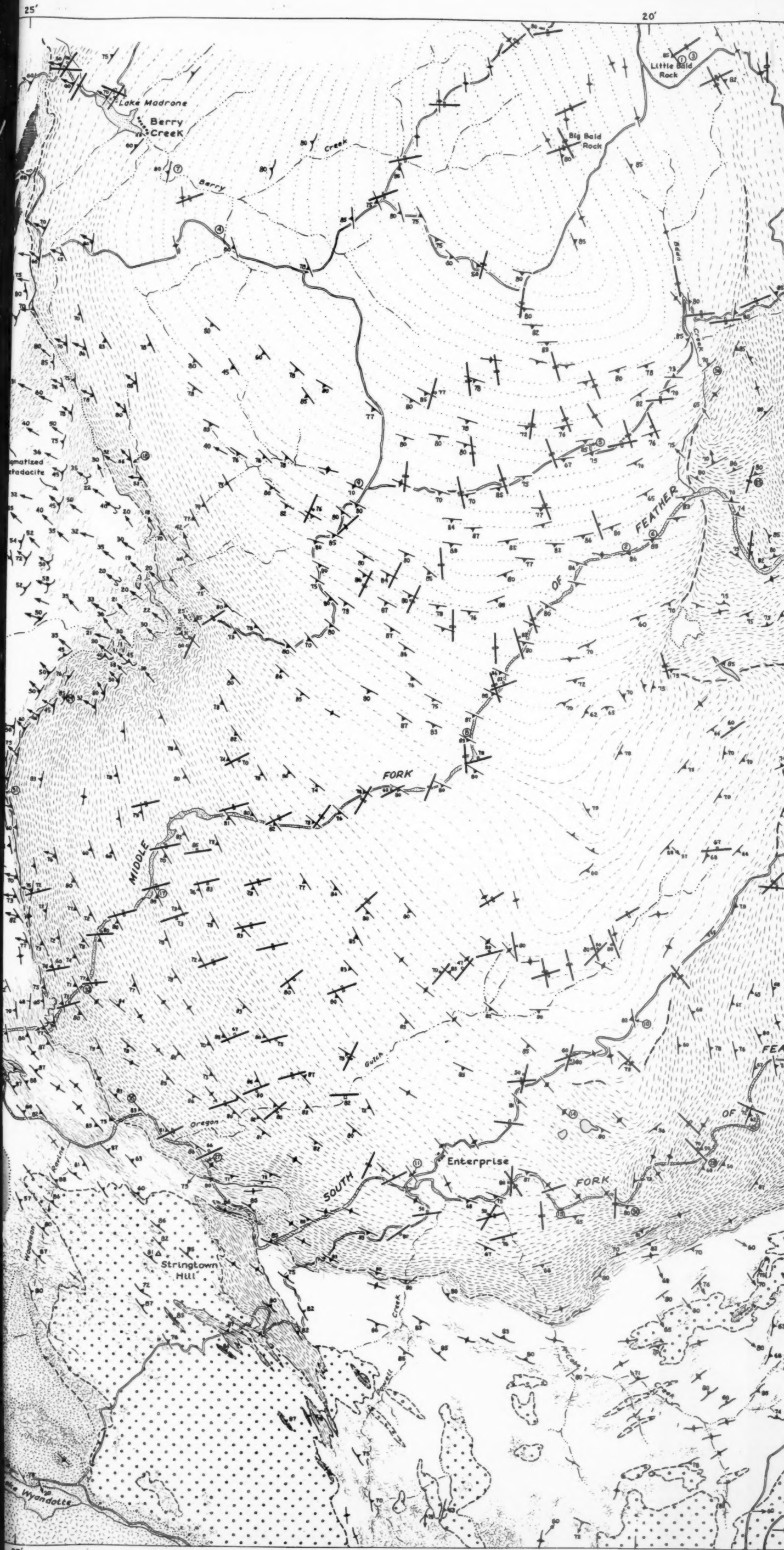
FEATHER RIVER

W.P.R.R.

Bidwell Bar Canyon Creek

Lake Wyandotte


Lake Madrone
Berry Creek
Berry Creek
Little Bald Rock
Big Bald Rock
Beers Creek
FEATHER
RIVER
FORK
MIDDLE
FORK
Oregon Gulch
SOUTH FORK
Enterprise
Stringtown Hill
Spanish Creek
Indian Creek
Rabbit Creek
Poverty Flat
Poverty Flat
Indian Valley
Indian Valley
Indian Valley



GEOLOGIC MAP OF BIDWELL BAR REGION, CALIFORNIA

121° 15′

37° 35′

37° 30′

121° 15′

Feather Falls

Fall River

RIVER

Swain Hill

Pine Mills Camp

Feather Falls

Run

Sucker

FEATHER RIVER

Old Forbestown

Sunset Hill

New Forbestown

Creek

Ravine

STRUCTURE SYMBOLS

Well located sharp contacts

Well exposed gradational contacts

Inferred contacts

Strike and dip of bedding

Strike and dip of schistosity, slip cleavage, and, in granitic rocks, of flow structure

(1) (2) (3)

Plunge of (1) crumple axes, (2) amphibole prisms or mineral streaks, and (3) cleavage prisms

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 66, PP. 9-44, 15 FIGS., 5 PLS. JANUARY 1955

TRONDHJEMITE BATHOLITH NEAR BIDWELL BAR, CALIFORNIA

BY ROBERT R. COMPTON

ABSTRACT

Bald Rock batholith, California, is one of a chain of small plutons that lies 20 miles west of the main Sierra Nevada pluton and parallels the N.15°W. structural trend of the northern Sierra. The country rocks are mainly late Paleozoic (?) basic volcanic rocks cut by small masses of basic plutonic rocks prior to late Jurassic time and folded and dynamothermally metamorphosed in the Nevadan orogeny. Foliation in the pre-batholith terrain was vertical, and its mineral assemblages were, with local anomalies, typical of greenschist facies. After regional metamorphism, gabbroic, dioritic, and granitic rocks were emplaced south and east of the site of Bald Rock batholith. The basic rocks of this sequence are in part magmatic, but the granitic bodies show gradational borders and internal structures that indicate static metasomatic growth, perhaps related to magmatic cores.

Bald Rock batholith was a forceful intrusion, as is shown by the abrupt concordant swing of bedding around it. However, local crosscutting relations show that about a fourth of its area at the exposed level was gained by other means. Large concentric outliers and a hull of injection migmatite suggest stoping. Inclusions are too scarce to account for the missing wall rocks, but gradational zoning of the batholith from a trondhjemite core through granodiorite to a heterogeneous tonalite rim suggests that basic stoped rock contaminated an originally trondhjemitic magma. This is supported by comparing a computed chemical composition of the granodiorite and tonalite with chemical analyses of two trondhjemites. The benign nature and pattern of flow structures shows that the different granitic rocks could not have been intruded separately. In contrast to the wall rock, plagioclase of the granitic rocks shows pronounced oscillatory zoning, indicating thorough reworking of the stoped rock. The reactions required for the changes were presumably exothermic, but the broad mixing shown here must have required additional heat.

Contact-metamorphic rocks of epidote amphibolite to possible pyroxene hornfels facies form an aureole that had an area nearly as great as the original kinetic intrusion. Squeezing and vertical stretching of these rocks is indicated by pebble and cyrstal lineation, most pronounced near the contact. Post-kinetic replacement veins of tonalite occur sporadically throughout the inner part of the aureole. The rise in anorthite content of plagioclase in metabasaltic rocks is gradual toward the batholith and sharp near the tonalite veins. The obvious fracture control of these high-temperature metamorphic structures indicates that heat addition to the aureole was largely through fluid transfer rather than simple thermal diffusion.

CONTENTS

TEXT

	Page
Introduction	10
Acknowledgments	11
Geologic setting	11
Principal rock units and their structural relations	11
Geologic history	12
Rocks of regional metamorphic terrane	13
General discussion	13
Nonmetamorphic structures	14
Original structures	14
General structural pattern	14
Metamorphic structures	14
Slip cleavage	14
Schistosity	15
Lineation	15
Origin of metamorphic structures	16
Petrography	16
Interpretation of regional terrane	17
Swedes Flat pluton and similar plutonic masses	18
General statement	18

	Page
Heterogeneous gabbroic and dioritic rocks	18
Occurrence	18
Structure	18
Petrography	19
Interpretation	20
Granitic rocks of Swedes Flat pluton	20
General	20
Structure	20
Petrography of tonalite and granodiorite	21
Petrography of inclusions	22
Petrography of granophyres	22
Emplacement of Swedes Flat granitic rocks	23
Granitic mass exposed along Big Creek	24
Bald Rock batholith	24
General statement	24
Structures	24
Relation to country-rock structures	24
Contact migmatites	25
Inclusions	26
Flow structures	26
Spatial relation between flow structure and rock type	27
Unconformable junctions of flow surfaces	27
Meaning of flow structures	28
Late dikes and irregular masses	28
Slip structures	29
Petrography	30
General	30
Gradational nature of rocks	30
Megascopic characteristics	31
Microscopic characteristics	31
Composition	32
Petrography of dike rocks	32
Emplacement of Bald Rock batholith	33
Interpretation of structures	33
Interpretation of compositional variations	34
Problems concerning contamination	36
Contact-metamorphic rocks	37
General statement	37
Structures	37
Bedding	37
Foliation and lineation	38
Randomly distributed migmatites	38
Late phyllonite sheets	38
Petrography of contact aureole	39
Textures of metabasaltic rocks	39
Mineralogy of metabasaltic rocks	39
Metadacites	40
Petrography of metasedimentary rocks	40
Petrography of randomly distributed migmatites	41
Retrograde or hydrothermal effects	42
Interpretation of contact aureole	42
Mineral-zoning and facies considerations	42
Mechanism of heat transfer to aureole	42
Structural effects of batholith on its country rocks	43
Summary of Nevadan plutonism	43
References cited	44

ILLUSTRATIONS

Figure		Page
1.	Geologic map of a part of northwestern Sierra Nevada	12
2.	Massive metadolerite cut by sheets of greenschist	15
3.	Schist after augite porphyry	15
4.	Lineation of slip cleavage prisms in schistose and semischistose metavolcanic rocks	16
5.	Swedes Flat tonalite	21
6.	Map showing details of contact migmatite	25
7.	Growth of Bald Rock batholith and adjacent plutons	29
8.	Crystallization of Bald Rock batholith as shown by flow structures	29
9.	Modes of thirty eight Bald Rock batholith rocks	30
10.	Spatial variation of plagioclase composition in batholith	33
11.	Variation diagram of Bald Rock batholith rocks	35
12.	Phyllonite sheets cutting layered hornfels and amphibolite	39
13.	Textural changes in massive metabasaltic rocks of regional terrane and contact aureole	40
14.	Distribution of plagioclase compositions in metabasaltic rocks of contact aureole	41
15.	Relation of linear structures to batholith	43

Plate		Facing page
1.	Geologic map of Bidwell Bar region, California	9
		Following page
2.	Hand specimens showing replacement migmatite and granitic rocks	22
3.	Replacement migmatites of Swedes Flat pluton	22
4.	Replacement migmatites in Bald Rock batholith aureole	22
5.	Replacement migmatites in Bald Rock batholith aureole	22

TABLE

Table		Page
1.	Mineral assemblages in rocks of regional terrane	17

Introduction

The metamorphosed volcanic and sedimentary rocks of the northwest part of the Sierra Nevada are cut by several small batholiths and plutons of granitic rocks. One of these, here called Bald Rock batholith, is well exposed east of Oroville where three forks of Feather River have cut deep, craggy canyons. Since 1947 the writer has studied this batholith and its country rocks, mapped 140 square miles at a scale of 1:24,000, and examined about 500 specimens petrographically. Meanwhile, Hietanen (1951) has described many aspects of the petrology of the area to the north. In view of overlap, this paper is restricted to a

description and explanation of features relating to Bald Rock batholith. The principal divisions cover the nature of the pre-batholith country rocks, the way the batholith was emplaced, and its endomorphic-exomorphic relations. Briefly, the evidence points to forceful intrusion of hot trondhjemitic magma into a terrane of greenschists and other low-grade rocks; large-scale contamination of the trondhjemite by metabasalt to form intermediate granitic rocks; and the formation of a high-grade aureole of amphibolite and replacement migmatite.

Acknowledgments

Thanks are due A. C. Waters and C. O. Hutton who advised the writer while he was working on parts of the problem as a student at Stanford University, and to W. I. Gardner and R. L. Gamer of the Bureau of Reclamation who kindly furnished preliminary copies of new topographic maps of the region and assisted in other ways. The Standard Oil Company of California's Stanford Fellowship in Geology helped finance the field work. Grants from Stanford University, in part from the Shell Grant for Fundamental Research, paid for some of the thin sections and the typing. A. Tihanravov and R. Laniz helped prepare materials for several of the figures. C. O. Hutton kindly read and criticized the manuscript.

Geologic Setting

Principal Rock Units and Their Structural Relations

About half the pre-Cretaceous rocks of the northern Sierra Nevada are granitic, and half are metamorphosed sedimentary and volcanic rocks; basic and ultrabasic plutonic igneous rocks are in the minority. The granitic rocks are divisible into two geographic units. The essentially continuous mass of the main Sierra Nevada pluton lies to the east; 20 miles west of it is a group of about a dozen closely spaced small batholiths and irregular plutons, one of which is Bald Rock batholith (Fig. 1). The country rock between the main Sierra pluton and the western group of granitic bodies consists chiefly of metasedimentary rocks, but the belt of rocks surrounding the western group is essentially basaltic, consisting of about 85 per cent metavolcanic rocks, 10 per cent slate, marble, and quartzite, and 5 per cent metaserpentine and metagabbro. The general alignment of the two metamorphic belts, as well as the bedding and foliation[1] within them, follows the Sierra Nevada trend of about N. 15° W. The irregular west border of the main Sierra pluton trends in the same general direction, and the western granitic masses, though they individually cut across this trend, are grouped in a similarly aligned belt.

The major, and best known, part of the western belt of igneous and metamorphic rocks is shown in Figure 1. Bedding and metamorphic foliation in this area are almost everywhere parallel or subparallel and dip steeply. Thus, at the scale used, the bedding trends are nearly independent of topography, and the map shows the pattern of country-rock structure in relation to the igneous masses. The four larger granitic masses in the northern half of the map, including Bucks and Merrimac plutons and Bald Rock batholith, are essentially concordant, though the detailed shape of their contracts shows that each has either stoped or replaced fair-sized chunks of country rock. The uniform regional trend of the country rocks bends so strikingly around these steep-sided masses, even with respect to their minor bulges, that it is reasonable to suggest that they were emplaced largely by forceful magmatic intrusion. However, the structural relationships of the the other igneous masses are quite different. The regional sweep of the country rocks is only moderately deflected around them. Where the author has studied these masses, most of their contacts are irregular and gradational. Large quantities of dark diorite and gabbro near their borders are distinctive. The contacts between these and the granitic rocks are complexly veined and gradational. A detailed study of the north end of Swedes Flat pluton suggests that a large part of these plutons is of replacement origin.

[1] The term "foliation" as used in this paper includes all platy structures except bedding, especially schistosity, gneissose structure, and metamorphic cleavage.

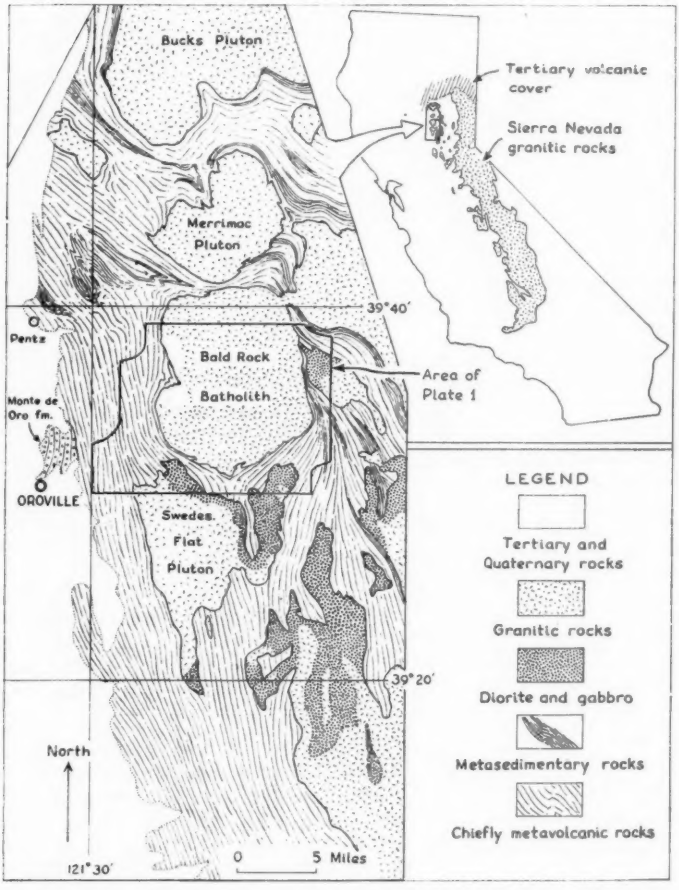


FIGURE 1.—GEOLOGIC MAP OF A PART OF NORTHWESTERN SIERRA NEVADA

Lined pattern shows general trends of bedding and foliation in metamorphic rocks; outline map of California indicates position of area with regard to main Sierra Nevada pluton. Northern third of map chiefly after Hietanen (1951, Pl. 1); southern third chiefly after Becker, Lindgren, and Turner (1895).

Geologic History

The solution of the geologic history of the Sierra Nevada region is badly impeded by the scarcity of fossils, the general destruction of critical sedimentary structures by metamorphism, the vast layers and lenses of volcanic rocks mixed with the sedimentary units, and the lithologic similarity of rocks of greatly different ages. No doubt careful field work will overcome these difficulties, but the literature available today scarcely alters the general geologic history set forth by G. F. Becker, H. W. Turner, and Waldemar Lindgren in their "Description of the Gold Belt", a supplement to the Gold Belt geologic folios (*see especially* H. W. Turner, 1898). With minor changes, their picture involves:

(1) Late Paleozoic sedimentation and volcanism in a geosyncline approximately coextensive with the Sierra Nevada; the principal rocks formed were shale and basic volcanic rocks, with subordinate limestone, chert, conglomerate, graywacke, and spilitic lavas.

(2) Probable Permian or early Triassic

orogeny, with emplacement of basic and ultrabasic plutonic rocks and perhaps some granitic rocks.

(3) Erosion and renewed marine deposition, during the late Triassic (?) and all but the latest Jurassic, of a suite of rocks similar to those of the late Paleozoic systems.

(4) Latest Jurassic (Nevadan) orogeny, with synkinematic intrusion of ultrabasic rocks and postkinematic emplacement of gabbro, diorite, and large quantities of granitic rocks, chiefly granodiorite. The postkinematic rocks may be early Cretaceous, since the oldest sediments overlying them are late Cretaceous.

The Colfax quadrangle (Lindgren, 1900), about 40 miles southeast of Oroville, offers evidence for all parts of this chronology. Geologic work in the area shown in Figure 1 offers little support for this sequence of events, though it in no way disproves it. Only two fossil localities are known, and, since they are far from the plutonic igneous masses, or rocks that strike into them, these igneous rocks can be dated only by lithologic similarity to known Nevadan igneous rocks.

Rocks of Regional Metamorphic Terrane

General Discussion

The metamorphic rocks mapped were originally a thick series of volcanic flows, pyroclastic beds, sediments, and minor volcanic and plutonic intrusions. All were subjected to low-grade regional metamorphism during the Nevadan orogeny and perhaps in part during a Permian orogeny. These rocks are part of a regional metamorphic terrane of considerable extent (Fig. 1). The Nevadan plutons emplaced into these low-grade rocks superimposed a distinct contact metamorphism on them.

South and west of Bald Rock batholith the country rocks are metavolcanic. On the east side of the batholith metavolcanic rocks are abundant, but metaserpentine and nontuffaceous metasedimentary rocks occur together in two extensive bands. By far the majority of the metavolcanic rocks were originally basalt or pyroxene andesite flows, dolerite flows and sills, and basic tuffs or tuffaceous sediments. Several small intrusive bodies of metarhyolite and metadacite occur in a chain extending north-south through Bidwell Bar. Metadacite, apparently originally in sills, is abundant near the west border of the batholith. A laccolith of actinolite-chlorite schist, originally a pyroxene porphyry, occurs 2 miles north of Bidwell Bar. Pre-Nevadan metadiorite and metatonalite migmatites and small metagabbro bodies occur west of the batholith, and a large concordant mass of metamorphosed granodiorite porphyry lies near its northeast border. These distinctive rock masses were mapped separately wherever possible. H. W. Turner called the metavolcanic rocks of the mapped area "Amphibolite" (1894; 1898) or "Greenstone schist series" (1896, p. 551) and "Hornblende series" (1896, p. 580). When the stratigraphic unity of these rocks is demonstrated by regional mapping, a formational name should be applied to them, but for this study they are referred to generally as the rocks of the regional terrane. H. W. Turner (1898) assigned the metasedimentary rocks that occur in the east side of the area to the Calaveras formation. Lindgren (1900) raised Calaveras to group rank when he subdivided Calaveras rocks into six formations. Though the metasedimentary rocks described in this paper probably belong to the Calaveras group, the writer uses their lithologic names only, since they have not been traced definitely to any one of Lindgren's units.

The only known Paleozoic fossil localities in the area of Figure 1 are in the metasedimentary rocks east and northeast of Pentz. The fossils collected here are Pennsylvanian or Permian (H. W. Turner, 1894, p. 446; 1898, p. 549). The only other means for dating rocks of the regional terrane is by projecting fossiliferous Calaveras rocks of the Colfax quadrangle along their strike to include the metasedimentary rocks that pass around the east side of Bald Rock batholith. This correlation is only approximate, but it seems reasonable. The Colfax rocks cannot be more closely dated than upper Paleozoic. The volcanic rocks are clearly interbedded with these sedimentary rocks in the east part of the area mapped, throughout the region to the north (Hietanen, 1951, p. 598), and in the northern part of the Colfax quadrangle (Ferguson and Gannett, 1932, p. 6–10). Thus, the metavolcanic rocks

in the east part of the area are probably late Paleozoic, while those of the west part are Pennsylvanian or Permian. There is no indication that the intervening metamorphic rocks are not essentially of intermediate age, though Triassic and Jurassic rocks may also occur here.

The only known Mesozoic fossil locality in the region is in the Monte de Oro rocks, 2 miles northeast of Oroville. Here, leaf-bearing slates suggest a Jurassic age. The associated rocks are chiefly tuff-breccias, lavas, and sandstones. They are notably less metamorphosed than the rocks of the regional terrane and appear faulted against them.

Nonmetamorphic Structures

Original structures.—Bedding of the volcanic and sedimentary rocks of the regional terrane can be recognized in many places, and the intrusive nature of dikes, sills, and laccoliths is, in some cases, determinable. That the gross layering is original rather than metamorphic is shown by the fact that the layers are several feet or tens of feet thick, that the rock textures are commonly well preserved, and that the changes in texture and composition at layer boundaries are distinct. The identification of original structures in the contact-metamorphic aureole is less reliable, for critical small-scale features are greatly modified. However, a gross layering parallel to near-by low-grade rocks is considered bedding. In all the metamorphic rocks, bedding is nearly parallel to planar metamorphic structures, so that it was only occasionally mapped separately.

General structural pattern.—The regional N. 15° W. bedding trend sharply splits south of Bald Rock batholith and swings in arcs of 70° around the southeast and southwest borders of the intrusion (Fig. 1). Toward the east and west, these arcs grade into the regional trend. The dip of the bedding is rarely less than 75° and is commonly near 90°. In part, changes in dip are caused by minor flexures, in other cases they are clearly related to tightly pinched chevron and isoclinal folds. In general, the folds observed in closure are small (several tens of feet) and plunge steeply. The absence of observable nonplunging folds may be due to the lack of large vertical exposures or to shearing on steep surfaces that strike about parallel to what may have been the fold axes. The forced intrusion of Bald Rock batholith may have produced just such slip effects. Dip patterns suggest more folds than can be proved; however, the only conceivable indications of large folds occur on the east side of the area where serpentine may core two major isoclinal folds. Though only a guess, the simplest picture consistent with the structural details and with the regional pattern of dated rock units is that of a tightly crumpled west-dipping homocline consisting, in the mapped area, of 10,000–25,000 feet of bedded rocks.

Metamorphic Structures

Slip cleavage.—Of the rocks of the regional terrane, about a third are schistose. The remainder are essentially granular and massive, though cut to a varying degree by a cleavage produced by slip on distinct surfaces. These slip surfaces are most commonly several inches apart; in the massive rocks they are several feet apart. Where they are closely spaced, the rocks grade into schists, though in many cases the schists are cut by slip cleavage at a low angle to their schistosity. The cleavage surfaces, where fresh, exhibit a sheen produced by tiny oriented actinolite needles or chlorite flakes; in more weathered outcrops, they are characterized by a reddish limonite film. Thin layers of fine-grained schist occur on many slip surfaces and are especially noticeable in the coarser metagabbros and metadiorites. Individual schist layers are of variable thickness, especially where the slip surface is irregular; these irregular layers commonly form a branching and joining system that slices the massive rock into separate lenses or flattened lensoid prisms (Fig. 2). Measurable displacements on slip cleavage occur in a number of places. Dark inclusions in a metarhyolite east of Bidwell Bar are offset as much as 6 inches along many closely spaced slip surfaces. A metadolerite dike 2 miles northeast of Forbestown is cut by about 12 slip surfaces, 6–10 inches apart, which have an aggregate displacement of at least 4 feet. The horizontal component of slip is large in all such cases, and the offset on each of a group of surfaces is in the same general direction.

Under the microscope, slip cleavage with no

megascopically visible schist layer appears as undulating, often discontinuous lines that distinctly transect the older fabric, including, in some cases, schistosity (Fig. 3). Fine, individual slips can be distinguished by small amounts of

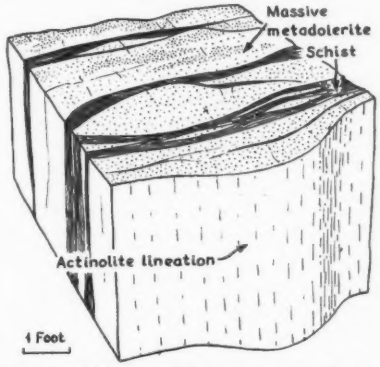


FIGURE 2.—MASSIVE METADOLERITE CUT BY SHEETS OF GREENSCHIST
Outcrop a mile northeast of Bidwell Bar.

leucoxene or limonite along them and by the way they displace grains. Where a few flakes of chlorite are oriented on them, the slips are more obvious, and, as a microlayer of schistose material is built up, the slip surfaces grade to megascopic schist layers.

Schistosity.—There are also more continuous bodies of somewhat coarser-grained schist. These have formed preferentially from tuffs, tuff-breccias, fine-grained sedimentary rocks, graywackes, pyroxene porphyries, and serpentinous rocks. The schist bodies are essentially giant lenses that interfinger with massive rocks along their strike. The heterogeneous pattern of schistose and massive rocks can best be pictured by adding the detailed pattern of Figure 2 to the regional pattern of Plate 1.

Under the microscope the schists show fine- and coarse-grained layers, some with better grain orientation than others. The schistosity is formed by subparallel actinolite, chlorite, or mica and, more rarely, by thin lenses of non-oriented minerals. Actinolite is common as lenticular grains whose ends are shredded out along the trace of the schistosity (Fig. 3). The tendency for schists to be cut by slip cleavage is especially noticeable in thin section.

Lineation.—The thin layers of schist on slip surfaces commonly show a parallelism of actinolite needles and, in some places, of elongate smears of light and dark minerals or of small crinkles. All three elements of lineation plunge steeply in most cases and lie essentially on the

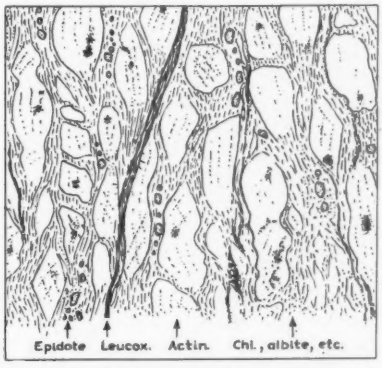


FIGURE 3.—SCHIST AFTER AUGITE PORPHYRY
Note leucoxene on slip cleavage. ×50.

slip-cleavage surfaces. Where they occur together in one outcrop, they may all plunge steeply, or the crinkles may be nearly at right angles to the steep mineral lineations, which are consistently parallel. The degree of development of the lineations is unequal from place to place, apparently owing to rock type. In one hand specimen, they may be clear cut on one slip surface but absent on the others. Possibly, actinolite lineation is present in the even-grained schists as well as on slip cleavage, but this is subtle enough to require detailed petrofabric study.

A fourth kind of lineation, probably more widespread than the others, consists of parallel prisms of schist or massive rock. These prisms are formed by the intersections of either two distinct sets of planar slip cleavage (Fig. 4) or of curved surfaces of slip cleavage (Fig. 2). The individual linear elements are elongate slabs with rhombic or lens-shaped cross sections. This structure was overlooked during most of the mapping because it is obvious only in outcrops with large, broken, vertical faces. It was later recognized and mapped in the southeast part of the area and in revisited parts of the western band of low-grade rocks. The structure is common in the regional terrane and generally has a steep plunge. Even in areas of poor exposure,

the more resistant rocks crop out as pointed slabs that indicate the steep-plunging lineation.

Origin of metamorphic structures.—The slip cleavage and the fine-grained schists and lineations associated with it are of displacement

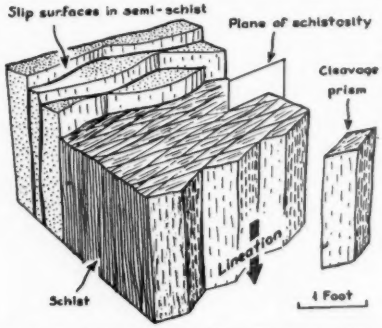


FIGURE 4.—LINEATION OF SLIP CLEAVAGE PRISMS IN SCHISTOSE AND SEMISCHISTOSE METAVOLCANIC ROCKS
4½ miles north of Bidwell Bar.

origin. The direction of movement on the slip surfaces has been largely strike slip. This is assured where the offsets can be measured and is strongly suggested by the consistently steep plunge of lineation formed by the lenticular pattern of joining or crossing slip cleavages. Some of the larger schist bodies may well be the end products of such displacement—that is, where the slip surfaces are fractions of a millimeter apart. However, schist bodies are commonly cut by slip cleavage, which allows that they may have originated in a different way.

In view of the local crosscutting relations of schistosity and slip cleavage, and considering the most obvious forces available, these structures may have formed during separate episodes. The first, the Nevadan orogeny, produced the larger, coarser-grained masses of schist and perhaps a part of the slip cleavage. The second, the intrusion of Bald Rock batholith, probably produced most of the slip cleavage, and particularly the cleavage that defines the steep plunging lineation of the rhombic and lensoid prisms. Because of the concordance of the batholith, the forces exerted on the regional terrane during its growth are, at any one place, largely parallel to the east-west forces of the Nevadan orogeny. Thus the structures produced during the two episodes are likely to be parallel or subparallel and therefore difficult to separate. More supporting evidence is provided by the linear structures of the contact aureole.

It is difficult to evaluate the structural effects of an earlier, perhaps Permian, orogeny on the terrane. The presence of thoroughly metamorphosed plutonic igneous rocks in the low-grade terrane indicates at least a period of pre-Nevada plutonism, which might have included dynamothermal metamorphism. If the present oriented structures of the terrane were formed in part by an earlier deformation, its forces must have been so directed as to produce essentially the same orientation pattern as that of the Nevadan orogeny. Furthermore, such a metamorphism must have been dynamically mild, for many of the rocks now show relic volcanic and sedimentary microstructures. It seems more likely that the present schistosity and slip cleavage are largely if not entirely Nevadan.

Petrography

Widespread palimpsest textures in the massive low-grade rocks indicate a variety of original rock types. Uralite pseudomorphs of pyroxene phenocrysts are generally present in rocks that were basalt or pyroxene andesite. Amygdules are well preserved in many places, and, in a few cases, even groundmass plagioclase laths retain their general shapes. Metadolerite, common in the metamorphic terrane, has ophitic arrangements of plagioclase and uralite. Metamorphosed gabbro, quartz-bearing diorite, and serpentine have relict granular textures. Even the fine-grained pyroclastic and epiclastic rocks, most commonly schistose, contain relict sand grains, small pebbles, or lapilli.

Though many gross relict features are recognizable, the newly grown metamorphic minerals tend to be fine-grained and intricately intermixed. In recrystallizing, plagioclase has commonly broken down into aggregates of small grains that together retain the shape of the original grain; it is never porphyroblastic. Where pyroxene has recrystallized to a single actinolite grain, its border is combed by acicular actinolite and clouded by leucoxene (Fig. 13,

left). Epidote is the only mineral that has locally grown to .3 mm grains; other minerals are seldom coarser-grained than .1 mm and in the fine-grained schists are commonly nearer .02 mm. Here and there, epidote and quartz, or chlorite and quartz, are segregated into veins, irregular clots, or gneissose bands, but otherwise the ion mobility of the reforming metamorphic materials was low.

In spite of the heterogeneity of the terrane, both in original rock type and in structural development, the present rocks are characterized by consistent mineral suites. Relict minerals are restricted to scattered grains of intermediate plagioclase, rare grains of pyroxene, and sphene. Acicular or uralitic amphibole is the most widespread and abundant metamorphic mineral and is probably actinolite in almost all cases, though $Z \wedge c$ is as high as 18° in some specimens, and birefringence is commonly as low as .017. Its color in Z may be evenly pale green or blue green with a splotchy pattern. Ny ranges from 1.63 to 1.65. A few grains of glaucophane were found in a metadacite, but they were isolated from actinolite. Plagioclase was determined by measurements of refractive indices in oils and, where possible, by measuring extinction angles without the universal stage. Compositions, checked with reasonable exactness in about half the rocks, are about An_4 to An_{12} in metabasalts and about An_4 to An_6 in metadacites and metasedimentary rocks. Epidote-clinozoisite, present in almost all the rocks, is generally iron rich but may be intricately mixed with or zoned to iron-poor epidote. Pumpelleyite occurs in some metabasalts and metadacites as fine-grained, granular aggregates in plagioclase or as minute prisms in quartz veinlets. Zoisite is common in some metagabbros. Chlorite, chiefly a pale-green prochlorite or perhaps ripidolite, is widespread but rarely abundant. Olive-green biotite is widespread but minor in all but the pelitic rocks, which are characterized by a pale, red-brown biotite. Dusty or finely granular leucoxene is present in most sections, commonly associated with limonite or scarce opaque grains. Talc and tremolite are essential constituents in metaserpentines, and muscovite is common in metasedimentary and acidic metavolcanic rocks. Quartz, almost everywhere present, is generally strained. Calcite is scarce, though widespread. Dolomite, pyrite, tourmaline, and fuchsite are scarce.

TABLE 1.—MINERAL ASSEMBLAGES IN ROCKS OF REGIONAL TERRANE

Approximate original rock type	Metamorphic minerals, in approximate order of abundance
Basalt or pyroxene andesite, gabbro, diorite	1. Actinolite-plagioclase (An_{4-12})-epidote 2. Actinolite-albite-epidote-chlorite
Quartz-bearing diorite	Albite-quartz-actinolite-chlorite-epidote
Dacite, rhyolite	Quartz-albite-muscovite-green biotite-epidote
Silty shale, graywacke	Quartz-pale red biotite-muscovite-albite
Tuffaceous or limy siltstone	Quartz-green biotite-actinolite-albite-epidote
Siltstone, sandstone, conglomerate	Quartz-pale red biotite-chlorite-albite-epidote
Serpentine	Talc-antigorite; talc-antigorite-tremolite

The metamorphic-mineral assemblages of the various original rock types are shown in Table 1. By far the most common mineral assemblage for both schistose and massive metabasaltic rocks is, in order of decreasing mineral per cents, actinolite-albite (An_{4-10})-epidote-chlorite-leucoxene, locally found with calcite and quartz.

Interpretation of Regional Terrane

The conditions of greenschist facies are indicated by the mineral assemblages of the metashales and metasiltstones (Table 1) and by the occasional assemblage actinolite-albite (An_4)-chlorite-epidote in metabasalt. The common occurrence of relict textures and structures and the generally fine-grained, fuzzy nature of the newly grown metamorphic materials indicate low-grade metamorphism. However, in rocks of basaltic composition the sporadic occurrence of plagioclase at least as calcic as An_{12}, and the scarcity of chlorite, suggests that

DETROIT PUBLIC LIBRARY

the conditions of the greenschist facies have been exceeded. The writer believes these differences cannot be explained on a basis of structural or textural control. Nor is it likely the oligoclase is relict from a previous period of metamorphism, for its composition is essentially constant within one sample, and there is apparently a complete range of compositions over the terrane. Finally, the An content shows no zoning relation to the contact aureoles of the granitic plutons. Perhaps this anomaly can be explained by considering the effect of H_2O and CO_2 content on the mineral assemblages of the metabasalts.

Calcium and perhaps silica tend to be removed from some basaltic rocks undergoing progressive low-grade metamorphism (Vogt, 1927; Sugi, 1931; F. J. Turner, 1935; Hutton, 1940; Saether, 1945). The resulting schists are enriched in albite and either chlorite or chlorite and stilpnomelane. As suggested by Eskola (1925, p. 91) and also by F. J. Turner (1933, p. 537–538), if calcium cannot migrate from such a system, and if there is insufficient CO_2 present to fix it as calcite, then abundant actinolite will form in greenschist facies. Presumably, temperatures in such systems are low, and, since the distances involved may be great, this removal of calcium must be affected chiefly by mobile aqueous solutions rather than any known kind of ion diffusion. Thus the richness of actinolite and the scarcity of chlorite and calcite in the regional terrane indicate a relatively dry metamorphism. Can the oligoclase in some of the rocks be explained by this dryness? If actinolite develops in place of chlorite, epidote is the only mineral that can hold the abundant aluminum released when intermediate plagioclase alters to albite. Epidote and actinolite are hydrous; if water is locally scarce, the only abundant anhydrous mineral, plagioclase, may be forced to hold more anorthite than is normal at a given temperature. This is only a suggestion; other reasons for thinking the regional terrane recrystallized under water-poor conditions are: (1) the rarity of mineral segregations, (2) the lack of porphyroblastic plagioclase, and (3) the general fine grain of the rocks. Rarity of relict minerals in all parts of the terrane argues against the idea.

If the terrane were indeed water poor during Nevadan metamorphism, are there any apparent causes for this condition? The predominance of massive lava flows and accessory intrusions over pyroclastic and epiclastic rocks suggests that it may have carried comparatively little water at burial. The pre-Nevadan period of plutonism may have partly dried the rocks before the Nevadan metamorphism. Until the mineral paragenesis of Sierra Nevada Paleozoic and Jurassic metavolcanic rocks can be compared regionally, these can be no more than suggestions.

Swedes Flat Pluton and Similar Plutonic Masses

General Statement

The north end of Swedes Flat pluton was studied in detail. The west end of a similar, smaller pluton on the east side of the mapped area was examined in a rather poor exposure. Both plutons are composite, consisting of heterogeneous gabbro and diorite and somewhat younger tonalite and granodiorite. Their more basic rocks are locally intruded and metamorphosed by Bald Rock batholith, and probably their granitic rocks are also somewhat older than the batholith. Like Bald Rock batholith, they are considered Nevadan and post-kinematic because they cut across the foliation of the regional terrane and because they are unaffected by regional metamorphism.

Heterogeneous Gabbroic and Dioritic Rocks

Occurrence.—Two heterogeneous masses of gabbroic and dioritic rocks, each covering about 4 square miles, occur at the north end of Swedes Flat pluton, and one of similar size underlies Swain Hill, on the east side of the mapped area. Countless dikes, veins, and smaller irregular bodies of dioritic rocks occur in the south-central part of the area, between Swedes Flat pluton and Bald Rock batholith. Only the larger ones appear on Plate 1.

Structure.—The larger masses are apparently aggregates or complexes of numerous small bodies emplaced at different times. These complexes are distinctly discordant to the bedding and foliation of the metamorphic rocks, and they are somewhat elongate parallel to the

strike of the country-rock structure. Most of the smaller bodies are tabular or lenslike, though some are lobate or even round in plan view. The exact contacts of the larger masses are difficult to locate because some of the rocks look much like the coarse amphibolite in which they may occur. Certainly, though, the contacts are irregular, especially where their general trend is athwart the country-rock structures. In some places dioritic rocks grade into amphibolite and basic hornfels as though the latter had simply recrystallized; in other places, particularly where the gabbroic and dioritic rocks are fine-grained, the contacts are sharp, and inclusions of hornfels occur in the igneous rock. The difficulty of mapping contacts can best be appreciated when the internal heterogeneity of the masses is seen. Irregular dolerite dikes and inclusions of either foliate or massive metavolcanic rocks are intermixed with coarser-grained gabbros and diorites of great textural variety. On Stringtown Hill, where one of the larger masses is well exposed, the inclusions are platy, several feet to tens of feet thick, and aligned with the regional trend of foliation as though they had been surrounded, but little moved, by the intrusions. The textural varieties of even the coarser rocks are so discontinuous from outcrop to outcrop their trends appear highly irregular; certainly they are not stratiform. Some rocks show a subparallel arrangement of light and dark layers ranging from an inch to several inches apart. Though some layering may be primary, most has been formed by saussuritization along fine parallel cracks. Where the fractures locally intersect, forming a boxwork of light bands, their secondary origin can hardly be doubted. At several places well within the masses, gabbro and dolerite show chilled borders to coarser rocks, and they also include fragments of these rocks. A widespread structural association is that of bodies of diorite pegmatite which occur in any or all other rock types and are especially concentrated near the periphery of the large complexes. These pegmatites range from one-crystal-thick veins to lobate bodies tens of feet across. The cores of the larger bodies consist of subradiate groups of hornblende crystals up to 18 inches long. Finally, all these rocks are cut by thin, widely spaced layers of fine- to medium-grained schist which parallel the contact of Bald Rock batholith and which probably were produced during its emplacement. The foliation symbols that appear within the gabbroic and dioritic areas on Plate 1 are mainly on these secondary schistose bands.

Petrography.—Uralitization and saussuritization have rendered most of the rocks gray green. The dolerites, however, are commonly fresh, and, locally even gabbro is fresh and black. Before deuteric (?) alteration, the rocks were chiefly hornblende-augite gabbro and dolerite, quartz-bearing augite-hornblende diorite, mesocratic hornblende tonalite, and diorite pegmatite. The augite is now largely uralitized, and the intermediate plagioclase is so altered the relative amounts of gabbro and diorite are difficult to ascertain. Apparently, the dioritic rocks form most of the Swedes Flat complexes, though gabbro may be the principal rock on Swains Hill. The only known occurrences of ultramafic rocks are two lenses of serpentinized pyroxenite, about 10 feet thick and 100 feet long, that occur near the north border of Swedes Flat pluton.

Microtextures range from granular to ophitic. Low-grade alteration has somewhat blurred the pyrogenic textures, typically producing shredded borders on the amphibole grains. Near Bald Rock batholith, the rocks have recrystallized to amphibolite in which only the general arrangement of dark and light minerals is retained.

The assemblages of pyrogenic minerals show consistent changes from the most basic gabbro to the most felsic diorites. The ratio of light to dark minerals ranges from about 1:2 in the gabbros to about 3:2 in the tonalite. Plagioclase, where fresh enough to be determined, is zoned from about An_{75} to An_{55} in the most basic gabbro, about An_{55} to An_{35} in the average diorite, and An_{35} to An_{25} in the diorite pegmatite. Quartz is absent from the gabbros but is present in the granular diorites, increasing in amount with the albite content of the plagioclase to a maximum of 15 per cent in one tonalite. Quartz occurs sporadically in the diorite pegmatites and generally forms less than 10 per cent of these rocks. The mafic minerals of the most basic gabbro are augite, pale-brown and gray-green hornblende, a little

hypersthene, and possibly scarce olivine. Augite and hornblende are the only mafic minerals of the other gabbros and quartz-poor diorites. The hornblende/augite ratio increases in the latter. The mafic minerals of the tonalite and diorite pegmatites are gray-green hornblende and some pale-brown biotite.

The deuteric, and possibly post-magmatic, minerals are dominated by uralitic actinolite which forms pseudomorphs after pyroxene and occurs as acicular overgrowths on pyrogenic hornblende. The uralite ranges from very pale green to distinct blue green in *Z* and is commonly irregularly tinted. Fine-grained opaque ore minerals, leucoxene, and possible sphene are common accessory products of uralitization. Chlorite and olive-green biotite are uncommon, and serpentine minerals rare. Abundant saussuritization products include epidote-clinozoisite, zoisite, calcite, sericite, and albite.

Interpretation.—The distribution patterns of the various rock types indicate that the large gabbro-diorite complexes were essentially foci for a myriad of small igneous intrusions and replacement or recrystallization bodies. The widespread occurrence of most of the rock types and the systematic change in their mineral parageneses point to a close relation among the various rocks, both within and between the large complexes. There is, however, no clear-cut univariant sequence of emplacement of the rocks. It can be concluded only that diorite pegmatites are the youngest rocks and that dolerite was intruded both before and after gabbro and diorite.

Chilled contacts and angular inclusions point to the igneous origin of much of the dolerite and at least some of the gabbro. The diorite pegmatite bodies, because of their gradational contacts and their lobate, irregular shapes, are probably largely of replacement or recrystallization origin. Little clear-cut evidence can be offered regarding the emplacement of the bulk of the dioritic and gabbroic rocks. However, because of their great textural heterogeneity, the presence of large, apparently undisturbed inclusions, and the local occurrence of gradational contacts, it is suggested they were formed by piecemeal intrusion combined with recrystallization and replacement of either metavolcanic rocks or earlier-formed diorite and gabbro. It is not possible to ascertain whether the widespread uralitization and saussuritization of the rocks is due to autometamorphism or to later alteration. Near Bald Rock batholith the rocks are essentially recrystallized in amphibolite facies, but, at distances of more than half a mile from the batholith, alteration is low grade and irregular. Much of the low-grade alteration in the large gabbro-diorite complexes at the north end of Swedes Flat Pluton may have been produced by the emplacement of Swedes Flat granitic rocks.

Granitic Rocks of Swedes Flat Pluton

General.—The principal granitic rocks of Swedes Flat pluton are tonalite and granodiorite which show, in the area mapped, a broad gradational contact zone with the metavolcanic country rocks. Most of these rocks are probably of replacement origin. Dikes and lobelike masses of granophyre along the northwest border of the pluton are apparently a local, late, intrusive phase. Both the granitic rocks and their contact zone are well exposed near Lake Wyandotte, where they were studied in detail. Brief reconnaissance a few miles south of the mapped area uncovered no data that conflict with those noted at the north end of the granitic mass. A separate small body of granitic rocks lies 6 miles east of Lake Wyandotte in the valley of New York Creek, but it is too poorly exposed to add anything critical to the study.

Structure.—Considering the Swedes Flat granitic mass exclusive of its western granophyre border, its gradational contact is striking. Dark hornfels and amphibolite give way to homogeneous granodiorite through a broad zone of intricately veined mixed rocks. This contact zone is only 100–1000 feet wide where the granodiorite contacts the heterogeneous masses of gabbro and diorite, but the gradations to metavolcanic rocks take place through distances as great as 3 miles. The dotted contact shown on Plate 1 indicates approximately where highly veined country rock passes into inclusion-charged granodiorite or tonalite. Only on the south border of the mapped area are the granitic rocks nearly free from inclusions and bunched mafic minerals.

Plate 2 illustrates the typical stages in the

development of massive, slightly clotted granodiorite from metabasaltic hornfels. Felsic material rarely first appears in the dark metamorphic rocks as separate porphyroblasts, but commonly as distinct veinlets and weblike clots. This apparent tendency for fracture control of the granitic material holds at all scales and is especially obvious in the more grossly diked rocks (Pl. 3, fig. 1). Even where the trends of these dikes are straight and parallel, their irregularly cusped borders and discontinuous nature make it difficult to picture them as dilation structures. For the most part, the dikes and veins are irregular and typically clotted in nodes or isolated star-shaped masses. They are locally oriented along the foliation of the metamorphic rocks (Pl. 3, fig. 2) but generally show no common orientation. Some of the individual beds of hornfels are considerably more veined than others; the finer-grained rocks are the preferred host.

The dark inclusions in the granitic rocks are not cored or zoned, and most show distinct, irregular contacts with their host. Almost all the inclusions are unoriented with regard to shape, and there is no other suggestion of planar-flow structures in the granitic mass. In the few cases where inclusions clearly show metamorphic foliation, this structure parallels the projection of country-rock foliation through the pluton. One of the most notable features is the lack of aplite or pegmatite dikes, either singly or in swarms.

The granophyres that form the west border of the pluton occur locally as thin veins and inclusion-charged masses, but also as thick dikes and lobe-shaped bodies that have sharp walls. The contact between the granophyre and the main granitic mass may be gradational, highly irregular, or both. It is exposed only well enough to indicate that it is not a simple contact. Some of the dikes of granophyre that extend into the country rocks to the west grade from medium-grained granophyre to porphyritic rocks with a very fine-grained groundmass. These porphyritic rocks resemble metamorphosed rhyolite or dacite, with which they occur locally. In the field, the metamorphic rocks can be distinguished only on the basis of their schistose borders or by the fact that they are cut by slip cleavage that parallels the regional trend. The granophyres have produced only a low-grade and narrow-contact aureole in the metavolcanic rocks in which they occur, in contrast to the other borders of Swedes Flat pluton.

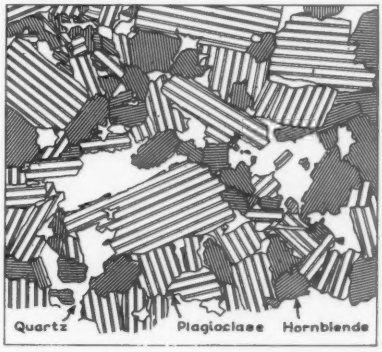


FIGURE 5.—SWEDES FLAT TONALITE
The quartz areas are almost all in optical continuity. 20×.

Petrography of tonalite and granodiorite.—In general, the mafic minerals of the granitic rocks tend to be clotted and their hypidiomorphic textures do not show flow alignment of feldspars or mafic grains (Pl. 2, fig. 4). The average grain sizes of the more homogeneous rocks range from about 2 to 4 mm, while those of the diked border zone range from 1 to 2 mm, with marked variability in individual dikes (Pl. 2, fig. 2). Under the microscope the textures show subhedral plagioclase grains and scattered mafic grains, both largely surrounded and partly replaced by intricately ramifying grains of quartz and minor alkali feldspar. In some cases spongelike quartz grains as much as 7 mm in diameter have preserved their optical continuity, and in most other cases quartz aggregates have been barely strained out of this sort of continuity (Fig. 5). The alkali feldspar is, at least in part, later than the quartz, for it locally cuts the quartz grains in thin irregular veinlets.

Quartz is unevenly distributed in the granitic rocks. On the average, it makes up about 8–15 per cent of the dikes and 20–40 per cent of the inclusion-free rocks. Plagioclase, which has yielded much space to quartz, is almost everywhere clouded by saussuritization; where it is

unaltered, it is progressively zoned from about An_{35} to An_{20}, with some cores as calcic as An_{45}. Oscillatory zoning of two to four rings is rare. Limited optical data obtained on the fine-grained alkali feldspar suggest that it is anorthoclase, though albite or microcline may also be present. Alkali feldspar forms as much as 10 per cent of the inclusion-free granodiorite, but it is rare or absent in granitic rocks of the contact zone. The mafic minerals are hornblende and biotite; the former is abundant everywhere. Like alkali feldspar, biotite is rare or absent in the border dikes and is important only in the nearly homogeneous granodiorite. Hornblende on the other hand makes up 20 per cent of the dikes and is locally concentrated near their borders (Pl. 2, fig. 2); it forms about 10 per cent of the nearly homogeneous granodiorite. Hornblende is gray green to pale brown in *Z*. Biotite is either in small olive-green flakes, replacing hornblende, or in larger red-brown flakes that locally rim hornblende. Epidote and apatite are the most widespread accessories. Fresh ore and sphene are scarce.

Petrography of inclusions.—Most of the inclusions in the dikes of the gradational border zone are sharply bounded and as dark and fine-grained as the wall rocks of the dikes in which they occur. Under the microscope they are granoblastic quartz-hornblende-oligoclase hornfels or andesine-hornblende amphibolite, about half of which is hornblende. Commonly they have a rim several millimeters wide in which grains grade from the finer-grained inclusion to the coarser dike. A few larger subhedral grains of hornblende or plagioclase may occur in or near this rim, and hornblende is quite commonly grouped in clots in it. The inclusions in the nearly homogeneous granitic rocks and some of the inclusions in the outlying dikes are coarser-grained and commonly lighter than the country rock. All the lighter-colored inclusions sectioned contain quartz in ramifying grains like those found in the granitic rocks, and many contain a little biotite and alkali feldspar. The darker inclusions have apparently been enriched in hornblende, and clots of hornblende in the nearly homogeneous granitic rocks are possibly relicts of such hornblendic inclusions.

Petrography of granophyres.—Megascopically, the granophyres are more leucocratic and somewhat finer-grained than the other granitic rocks. The finest-grained granophyres are indistinguishable from unsheared metarhyolite of the low-grade terrane. Under the microscope, the coarser granophyres show prominent micrographic and granophyric quartz-albite intergrowths intermixed with 2-mm grains of both minerals. In the granophyre which occurs as dikes along the west border of Swedes Flat pluton, the intergrowths are fine-grained and crudely spherulitic. Irregular miarolytic cavities, partly filled with euhedral quartz and epidote, are common in some of the finer-grained rocks. In coarser granophyres, the micrographic texture seems transitional to the

---

PLATE 2.—HAND SPECIMENS SHOWING REPLACEMENT MIGMATITE AND GRANITIC ROCKS

FIGURE 1.—METABASALTIC HORNFELS WITH DISCONTINUOUS QUARTZ-ANDESINE-HORNBLENDE VEINS 1½ miles northwest of Lake Wyandotte.

FIGURE 2.—NODES OF MEDIUM-GRAINED TONALITE IN REPLACEMENT MIGMATITE

Note hornblende clots (dark), below and right of center. 1 mile west of Lake Wyandotte.

FIGURE 3.—MASSIVE TONALITE FILLED WITH RELICS OF AMPHIBOLITE

1 mile north of Lake Wyandotte.

FIGURE 4.—GRANODIORITE WITH SMALL INCLUSIONS AND BUNCHED MAFIC MINERALS

On road, half a mile west of Lake Wyandotte.

PLATE 3.—REPLACEMENT MIGMATITES OF SWEDES FLAT PLUTON

FIGURE 1.—TONALITE DIKES WITH TYPICALLY CUSPED BORDERS

Note sharp borders and the few faded inclusions. Woodman Ravine, 1½ miles north of Lake Wyandotte

FIGURE 2.—CONTROL OF TONALITE VEINS BY COUNTRY ROCK FOLIATION

FIGURE 1

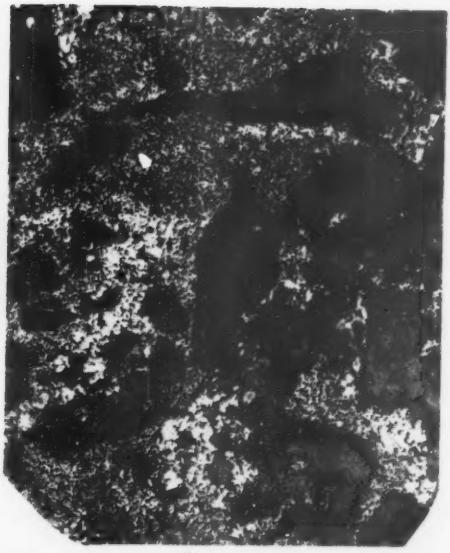


FIGURE 2

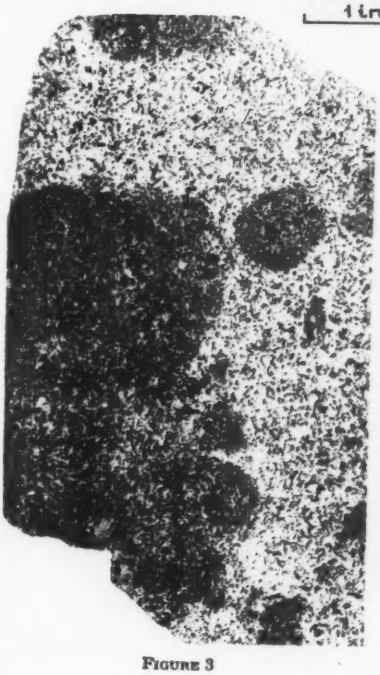


FIGURE 3

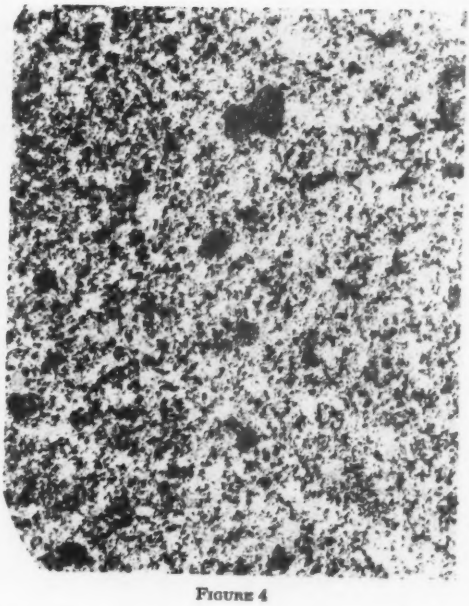


FIGURE 4

HAND SPECIMENS SHOWING REPLACEMENT MIGMATITE AND GRANITIC ROCKS

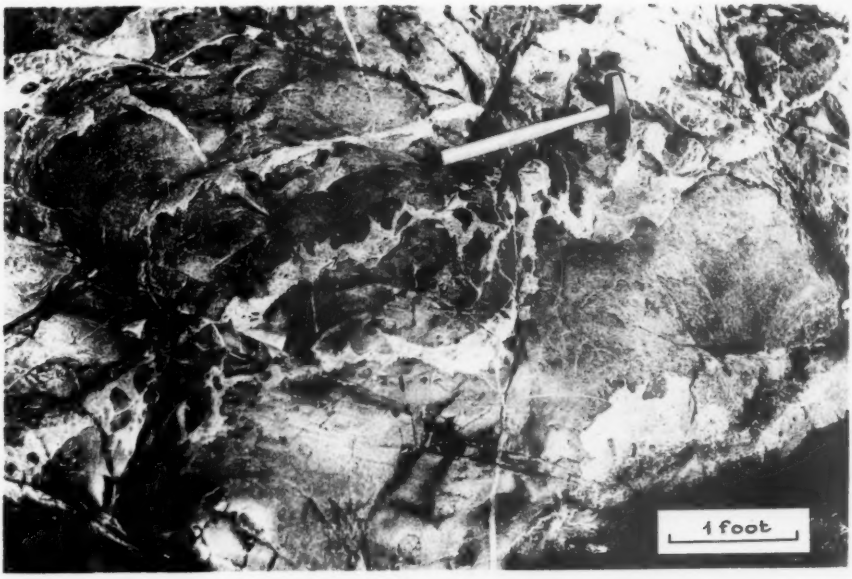


FIGURE 1

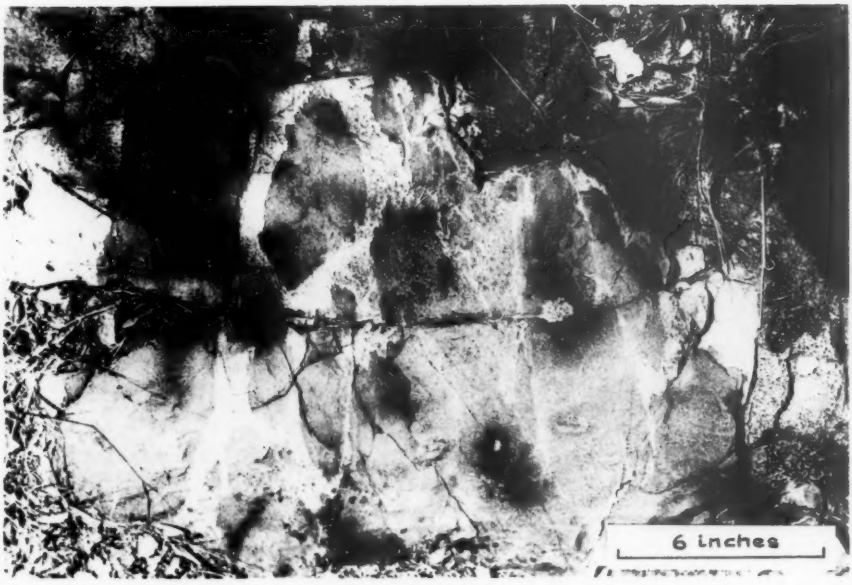


FIGURE 2

REPLACEMENT MIGMATITES OF SWEDES FLAT PLUTON

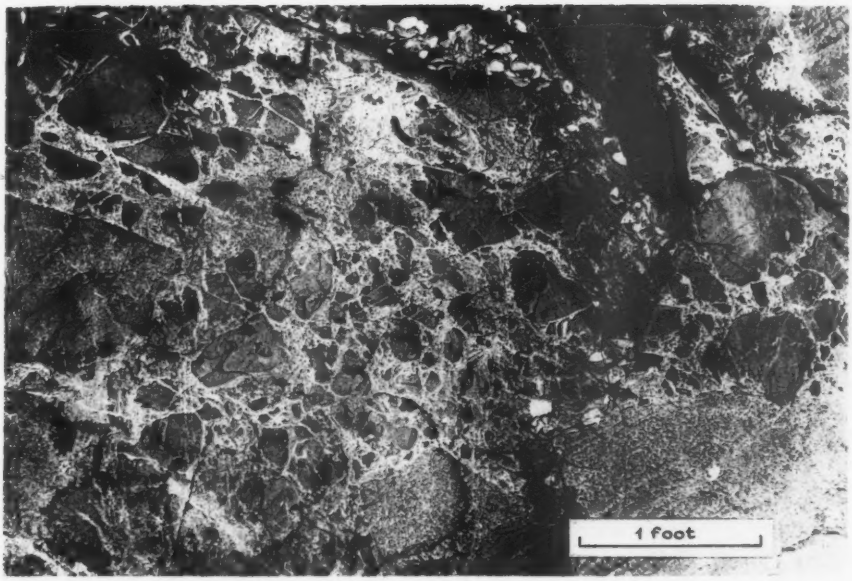


FIGURE 1

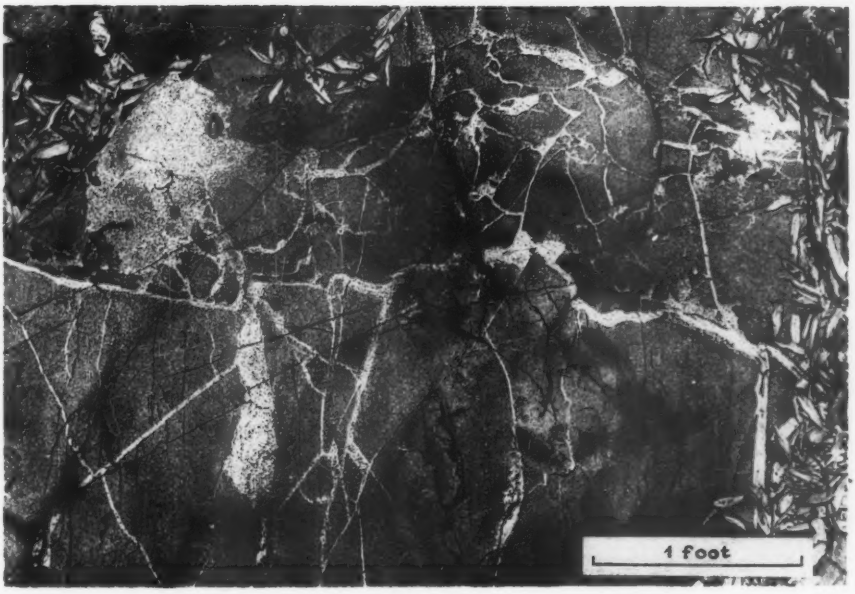


FIGURE 2

REPLACEMENT MIGMATITES IN BALD ROCK BATHOLITH AUREOLE

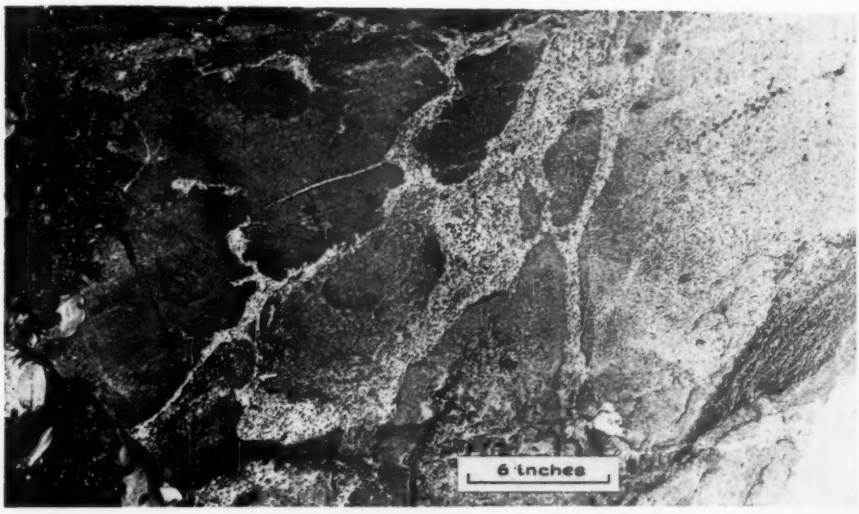


FIGURE 1

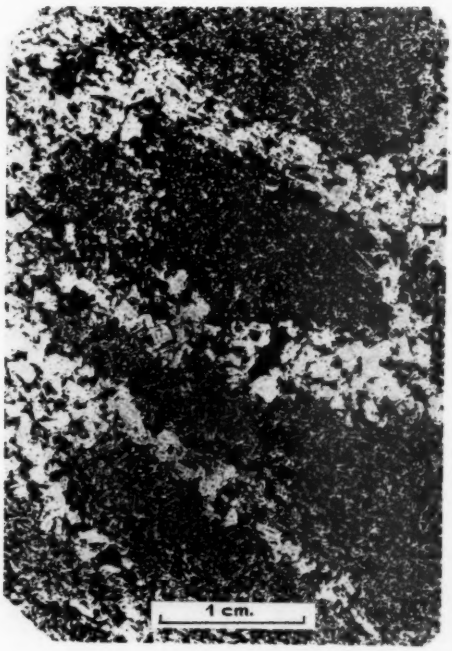


FIGURE 2

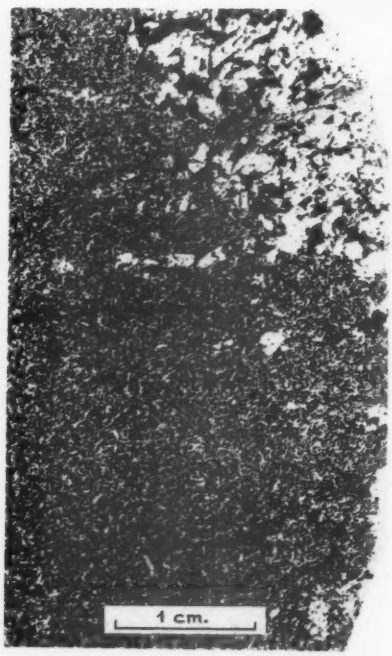


FIGURE 3

REPLACEMENT MIGMATITES IN BALD ROCK BATHOLITH AUREOLE

coarse quartz-plagioclase mixtures of the tonalite.

Quartz makes up 50–65 per cent of the rocks, and albite all but a few per cent of the remainder. A little orthoclase, olive-green biotite, and iron ore are the only other primary constituents. Epidote, sericite, and leucoxene are of regular and widespread occurrence and are probably deuteric. The mineral assemblage strongly suggests low-temperature crystallization from a water-rich magma or other fluid. Only one small pegmatite pod was found associated with the granophyre.

Emplacement of Swedes Flat granitic rocks.—A recrystallization and replacement origin for the veins and dikes of granitic rocks is indicated by their discontinuous, irregular, and weblike nature. The regular, gradational changes from the incipient webbing of these mixed rocks to inclusion-charged granitic rocks and then to granitic rocks with only vague hornblende clots suggest that the granitic rocks have passed through these stages in their development. A replacement origin for the homogeneous rocks is supported by their lack of flow structures or fracture pattern and by the fact that the foliate inclusions fit the projection of country-rock foliation through the pluton rather than any reasonable movement picture within it. Perhaps the most compelling evidence for static growth of the north end of the pluton is that the intricately diked border zone is nowhere smeared out, extended, or otherwise deformed. The preservation of optical continuity in the spongelike quartz grains indicates that at least the later stages of granitic emplacement were static.

Judging from the petrography of the veins and dikes, the first step in the granitization of the metavolcanic rocks was recrystallization along fractures with increase of plagioclase relative to hornblende, the latter tending to clot near the borders of the veins. As recrystallization progressed, soda, silica, and a little potash were added at the expense of the other constituents. Late in the development of the rocks, abundant silica was added, alkalies were available, and plagioclase was saussuritized. Perhaps the granophyres were formed during this last low-temperature stage, as mobile concentrations of silica-alkali pore fluids.

The repeated fracture control of the granitic veins and dikes indicates that most of the granitization must have been produced by fluids moving in open channels. The source of these fluids was probably a magma that formed the core of Swedes Flat pluton. This magma would have lain south of, and perhaps at a lower level than, the rocks described here. A pressing problem is the final resting place of the large quantity of calcium, magnesium, and iron removed by the granitization. If only the area

---

PLATE 4.—REPLACEMENT MIGMATITES IN BALD ROCK BATHOLITH AUREOLE

FIGURE 1.—REPLACEMENT BRECCIA BORDERING SMALL TONALITE POD

Powell Creek, 2 miles south of the Bald Rock batholith contact.

FIGURE 2.—HORNFELS WITH TONALITE VEINS AND NODES

Note stages of formation of inclusion-bearing nodes. Powell creek, 1½ miles south of batholith contact.

PLATE 5.—REPLACEMENT MIGMATITES IN BALD ROCK BATHOLITH AUREOLE

FIGURE 1.—VEINS OF TONALITE CUTTING PORPHYROBLASTIC HORNFELS WHICH GRADES TO TONALITE (RIGHT)

Powell Creek, 1½ miles south of batholith contact.

FIGURE 2.—THIN SECTION OF TONALITE VEINS ONE SHOWING FLOW STRUCTURE (TOP)

Degree of sharpness of veins typical. Note coarse hornblende locally bordering veins. McCabe Creek, 1½ miles from batholith contact. Unpolarized light.

FIGURE 3.—THIN SECTION OF TONALITE VEINS AND PORPHYROBLASTS IN AMPHIBOLITE

Dark streaks in hornfels are coarse-grained hornblende. McCabe Creek, 1 mile from batholith contact. Unpolarized light.

most affected by replacement—that is, the diked and inclusion-charged border zone of the pluton—is considered, it can be shown that about a third of the rocks are granitic. The loss of basic materials required for their emplacement is sufficient to increase the hornblende content of a 1000-foot border zone of metavolcanic rocks from about 55 per cent to at least 80 per cent by volume. The small hornblende clots that border many of the granitic veins and dikes indicate a tendency for such enrichment, but they account for very little of the addition required. Furthermore, the hornfels and amphibolite immediately adjoining the pluton are essentially the same as those occurring elsewhere. The gabbroic and dioritic rocks that border parts of the pluton cannot be the sort of basic byproduct required because their average composition is about the same as that of the metavolcanic rocks in which they occur. If the basic materials were moved ahead of the growing pluton, they must have escaped upward rather than outward. Perhaps they formed mafic gabbroic and dioritic rocks in the metavolcanic terrane that overlay the pluton.

Granitic Mass Exposed along Big Creek

The small rather poorly exposed tonalite and granodiorite pluton east of Bald Rock batholith was not studied exhaustively. It shows some structural features like those of Swedes Flat pluton and others like those of the magmatic Bald Rock batholith. Like Swedes Flat pluton, it is roughly aligned with the trend of the country rocks, and it is bordered on the north by a mass of heterogeneous gabbroic and dioritic rocks. Between the basic and granitic parts of the pluton is a veined and diked zone strikingly similar to that of Swedes Flat pluton, though much narrower. Within this border zone the petrography of the two plutons is similar. Inward from the border zone, however, the homogeneous granodiorite of the pluton differs from that of Swedes Flat pluton in that it has a planar flow structure and contains few inclusions or hornblende clots. Furthermore, its west contact with metamorphic rocks is sharp and has blocky re-entrants, in no way suggesting metasomatic emplacement. Under the microscope, the granodiorite shows a normal hypidiomorphic granular texture, distinct oscillatory zoning in plagioclase, and rare deuteric alteration.

Bald Rock Batholith

General Statement

Bald Rock batholith is a trondhjemite-granodiorite-tonalite intrusion about 80 square miles in area. Hietanen (1951) mapped and described its north end, and the remaining part was mapped and studied in detail by the writer. It is named after the large exfoliation domes in the north-central part of the intrusion.

Structures

Relation to country-rock structures.—Considering its squarish shape, Bald Rock batholith is remarkably concordant to the bedding and foliation of the country rocks. The regional N. 15° W. trend of bedding is split into a stubby Y at the south end of the batholith, and from there bedding swings evenly around its east and west sides. At its north end, Hietanen has mapped a striking east-west swing of metamorphic units essentially parallel to the batholith contact, but at about right angles to the normal Sierra Nevada trend (Fig. 1).

In detail, the contact is generally discordant. Its crosscutting effects range from small dikes and blocky migmatites to large-scale intrusive relations. Along the southwest border, large projections cut abruptly for a quarter of a mile or more across the layering of the metamorphic rocks. A more important, though less obvious, relation is the tendency for the batholith to cut the country rock at a low but consistent angle. This is particularly clear along the southeast border, where the average angle of discordance is about 5°. Similar relations are shown on Hietanen's map. Since bedding, foliation, and the outward dip of the batholith contact are steep almost everywhere, Plate 1 gives a reliable three-dimensional picture of the structure. If bedding is projected across the contact, starting at the southern apex, and continued subparallel to the trends in the nearest part of the contact aureole, it is seen that large slices of country rock are missing along the east and west borders of the intrusion. These slices have

the shape of crescents, thickest at the eastern and western extremities of the batholith. The eastern slice is at least 1½ miles thick near Feather Falls village. The greatest thickness of the west slice is about a mile, just east of Bidwell Bar.

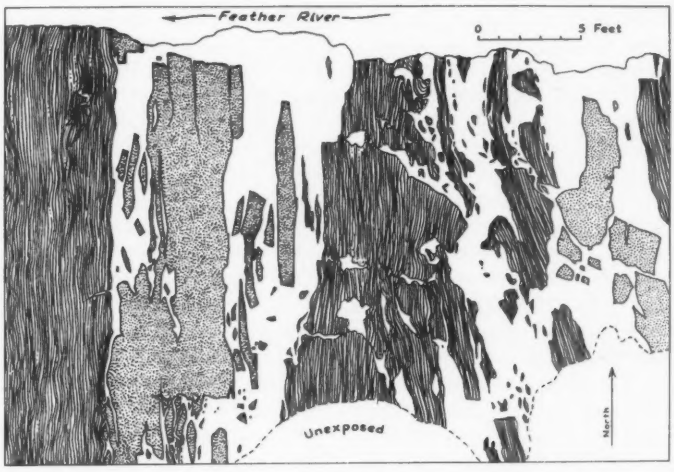


FIGURE 6.—MAP SHOWING DETAILS OF CONTACT MIGMATITE

Granitic rocks unpatterned. Some small dikes diagrammatic. Patterns in metamorphic rock show structure and extent of different layers. Broad zone of migmatites east of Bidwell Bar, ⅓ mile from batholith contact.

The structural relation of the batholith to its country rocks indicates two basic mechanisms of emplacement: (1) forceful intrusion resulting in strong outward forcing of the country rocks; and (2) lateral growth to the east and west by stoping, assimilating, or granitizing the country rocks. Judging from the approximate projection of country rock units through the batholith, (2) must account for about a quarter of the granitic rocks at the present level of exposure.

Contact migmatites.—All intermixed metamorphic and granitic rocks outside the main contacts of the batholith and its outliers are here called migmatite. The dikes of much of the migmatite differ only in size from the several granitic outliers. The outliers are sharply bounded bodies whose contacts can be traced at the map scale used; the migmatite dikes must be shown diagrammatically. To determine the significance of the migmatites, both their concentration and alignments were traced throughout the contact aureole. Two general types were discerned: contact migmatites, largely of injection origin, that occur immediately adjacent to the contact; and randomly distributed migmatites, largely of replacement origin, some of which occur as far as 2½ miles from the batholith.

The zone of contact migmatites is only a few feet to a few tens of feet wide where country-rock foliation is parallel to the contact; but it is as much as a quarter of a mile wide where the contact is sharply discordant to foliation. One exception is at the concordant western apex of the batholith, just east of Bidwell Bar. Here the migmatites are 4000 feet wide and lie between the batholith and a large outlier.

Most of the granitic parts of the contact migmatites occur as sharply bordered dikes that generally parallel foliation, but locally cross it. Dark angular inclusions are present in most dikes, and many of these have clearly been moved. Locally the dikes grade to bands of contact breccia (Fig. 6). Planar-flow structures that involve both mineral grains and inclusions are locally distinct; however, only those layers that have apparently been squeezed and sheared out, presumably by shouldering of the batholith, are strikingly gneissose, and they are characterized by tightly folded and lens-

shaped relics of granitic rock. They are far less abundant than the massive dikes, into which they grade. Intermixed with the clean, angular inclusion-charged dikes are discontinuous lenses, blebs, and irregular veins of granitic material that locally grade into porphyroblastic hornfels. These smaller features are primarily of replacement origin. They are widespread, but generally of much less bulk than the intrusive dikes, which in most cases cut them. Occasional aplite and pegmatite dikes cut at all angles across the migmatites.

Replacement or recrystallization migmatites are volumetrically important only along the northwest contact, between Lake Madrone and Canyon Creek. Here, metadacite and metarhyolite are common and apparently grade into light-colored bands of foliate migmatites that have a vague gneissose and veined structure. Though the area is not well exposed, the general relations are probably reliable, particularly the cutting off of the migmatized bands by the batholith (Pl. 1). Hietanen's (1951, p. 593) interpretation of the migmatized border in the northern half of this sector is essentially the same as the writer's, although she placed the batholith-migmatite contact farther east. This is the only place along the contact where the contact-migmatite zone appears gradational to the batholith rocks.

Judging from the repeated evidence of dilation, shattering, and inclusion movement in the dikes of the contact migmatites, most of them represent injections from the batholith. The slabs and fragments of metamorphic rock adjacent to the batholith must have been almost floating into it when the migmatite dikes were mobile. Several of such nearly severed masses are large enough to appear on Plate 1. Indeed, the nature of the contact migmatites suggests that piecemeal stoping was a very likely mechanism of batholith growth. If stoping takes place most easily where country-rock foliation is parallel to the contact, the pronounced eastward and westward extension of the intrusion is explained. Judging from the thick band of migmatite east of Bidwell Bar, the intrusion of large concordant outliers may have greatly increased the effect of a stoping mechanism. Metasomatism was clearly an effective, though in most cases a minor, process of migmatitization. It probably has little to do with the striking easterly and westerly extensions of the intrusion, for there is no apparent reason why metasomatism should have been more effective in these places than at the south apex.

Inclusions.—Though the contact migmatites suggest stoping, dark, easily recognized inclusions are rare in the batholith. They are distinctly concentrated within about a quarter of a mile of the contact and in a linear zone east of the center of the intrusion. Even here, however, they make up at most 2 per cent of the rock. Where inclusions are clustered, neighboring fragments differ greatly in color, grain size, and degree of foliation. Adjacent foliate inclusions may or may not be structurally aligned. The inclusions range from sharp, dark fragments to ghostlike masses that differ only slightly from the surrounding granitic rock. Some are zoned to hazy borders, others have faded evenly. Within one outcrop there may be a variety of inclusions indicating either that they represent several source beds, or else their post-inclusion histories have been different and they have been swept together by chance.

The scarcity of inclusions in Bald Rock batholith seems to negate the mechanism of piecemeal stoping. The only way the two can here be reconciled is to adopt the hypothesis that the dikes of the contact migmatites solidified first, and the adjacent part of the batholith was sufficiently fluid and reactive to dispose of most of its inclusions. In other words, assimilation or granitization of inclusions was somewhat faster than the inward growth of essentially crystalline granitic rock. The evidence for assimilation (using the term in its broadest sense) and for the idea that the batholith grew as a very mobile liquid mass lies largely in the relation between flow structures and the distribution of rock types within the batholith.

Flow structures.—Planar parellelism of biotite and hornblende can be detected at almost every outcrop of the batholith; inclusions and very scarce schlieren are oriented parallel to the crystals. Here and there, hornblende prisms show a steep lineation in the flow surfaces, but this structure is rarely distinct. Plagioclase, where tabular on {010}, accentuates the planar structure, particularly the tendency of the rocks to split along it. The structure must have

formed by benign and approximate flow orientation of well-formed crystals. It is strikingly gneissose only where protoclasis has smeared out the mafic minerals. The flow structure is more obvious near the contact because mafic minerals and inclusions are here most abundant, but its actual perfection, grain to grain, does not vary greatly from the rim to the core of the intrusion. The only exception is the protoclastic rocks, which are essentially restricted to the border zone. The flow structure can be mapped readily except for the few places where it is swirled or shows unconformable junctions within a small area of outcrops.

Spatial relation between flow structure and rock type.—The interrelations between the flow-structure pattern and the rock-type pattern is especially critical in determining the batholith's development. Therefore, the gross distribution of the granitic rocks must be outlined here; details of their variations will be described in the section on Petrography. On the combined geologic and structural map (Pl. 1), each flow symbol represents several readings; no symbols have been abstracted or changed from the field sheets. The red lines of the batholith rock patterns show the probable trends of flow surfaces, as projected between the symbols. They also show, by their spacing, the approximate distribution of rock types. The granitic rocks form a crudely concentric zonal pattern, the most leucocratic at the core. Tonalite, the most mafic rock, forms a nearly continuous border zone and a patchy north-south band that lies just east of the center line of the intrusion. The border zone is considerably broader in some places than in others. Significantly, the areas of tonalite coincide with those in which inclusions are most abundant. Tonalite grades into granodiorite, which makes up most of the east half of the intrusion and forms a broad crescent-shaped border in its southwest sector. The west half of the intrusion is dominated by an unbroken central mass of trondhjemite which is gradational to granodiorite along its borders. Toward the north, the trondhjemite becomes more leucocratic, locally containing only 2 per cent of biotite. Trondhjemite with less than 5 per cent of mafic minerals is here called leucotrondhjemite, and this rock makes up the core of the batholith. It grades evenly and broadly to the more mafic rocks of the west border, but its junction with granodiorite to the east is abrupt and locally intrusive.

Except for a few places along the east border of the trondhjemite mass, the flow structure passes conformably from one rock type to its neighbor. In some places the flow surfaces parallel the gradational boundaries between rock types, in other cases they cut across them at large angles. The overall pattern and the continuity of the flow surfaces are striking. Flow structures near the contact are everywhere parallel to it. Within 1–2 miles from the east and west contacts the flow surfaces follow the same trend for great distances. Near the southeast border the situation is different; the structures curve rapidly northeast away from the contact until they coincide with the northward trend of the east border. All the flow surfaces of the east half of the batholith impinge on the southeast contact zone. In the west half, a similar relation occurs along the west-central contact. The surfaces become more clearly concentric in the trondhjemite and are arranged elliptically around the leucotrondhjemite core. Notably, the flow structures dip as steeply at the core as anywhere in the intrusion. Besides the core, the only distinct closure in the flow surfaces is that of a tightly pinched syncline—a structure resembling a keel—which lies about 3 miles southeast of the core. Because this keel consists of tonalite and is charged with inclusions along its axis, it probably represents the base of a former roof pendant.

Unconformable junctions of flow surfaces.—There are several extensive, unconformable junctions of the flow surfaces that are probably local intrusive contacts. These are mapped as heavy black dashed lines. The most clearcut of these extends, with two major breaks, along the approximate north-south center line of the batholith. It forms a part of the boundary between trondhjemitic rocks to the west and more basic rocks to the east. An angular junction between trondhjemite and granodiorite occurs in the south-central part of the batholith, and a possible intrusive junction in granodiorite occurs 2 miles northwest of Feather Falls village. Less extensive angular junctions occur here and there where flow surfaces that are parallel for long distances locally cut each other at low

angles. Some of these were formed by late slip on surfaces nearly parallel to the flow layers; others are probably local unconformities formed by intrusion, for they show no effects of granulation. Near the tonalite keel, there were not enough outcrops to trace out flow directions in detail. Some of the swirls might be shown as angular junctions, and vice versa. This is the only generally complicated part of the flow-structure pattern. The marked small-scale swirling around the projection of country rock on the west border is more readily interpreted. It has apparently been formed by a southwestward squeezing from the trondhjemite core, when the border rocks were largely crystalline. This is indicated by the parallel crumpling of the country rocks and migmatite, and by protoclastic structures and textures in both the migmatites and batholith rocks. Similar though less well-marked structures occur locally on the southeast border of the intrusion.

Meaning of flow structures.—The overall continuity of the flow-structure pattern, and its independence of most rock-type boundaries, indicates that the different rocks are not the products of multiple intrusion; the flow structures must have formed by some process integral to the batholith as a whole. The manner in which the batholith flow layers impinge on its discordant contacts makes it difficult if not impossible to explain the pattern by the usual picture of upward welling and doming of a viscous, semicrystalline mush. Could the structures have risen by some post-intrusive mechanism? Possibly they represent stages of inward crystallization of the intrusion after it had extended its east and west boundaries as far as possible. Crystal orientation was then affected by continuous movement against progressively less mobile material. The progress of this interface of immobility began at the contact, to which the surfaces are everywhere parallel, and it grew inward most rapidly in the extended east and west sectors. The west half of the batholith was the focus of the late mobile magma, with leucotrondhjemite forming the last magmatic core. The stages of batholith growth and crystallization are summarized in Figures 7 and 8. Most of the angular unconformities represent surfaces of erosion between the semicrystalline border and the residual mobile magma. Protoclasis and crumpling were produced by late outward forcing of the magmatic core on those parts of the border mobile enough to deform and recrystallize under stress.

Late dikes and irregular masses.—The bipartite nature of the batholith, which apparently resulted from localization of the late trondhjemite magma in its west half, is emphasized by the kinds and patterns of late intrusive and replacement bodies—in the west half thousands of thin aplite and pegmatite dikes, in the east half several large dikes and pipes of aplite and microgranitic rocks. Dark dikes are rare in both parts of the batholith. The aplite and pegmatite dikes of the west half are concentrated along the axis of the trondhjemite mass; here their trends parallel the axis. Almost all the dikes are vertical and trend at about right angles to the flow structure. Thus they fan out on the flanks of the trondhjemite mass and, in the granodiorite and tonalite, lie at about right angles to the near-by batholith contact. The dikes are generally an inch to a few inches thick, have straight parallel contacts, and may consist of either or both aplite and pegmatite. They vary considerably in abundance. A symbol on Plate 1 indicates anything from several hundred to only three subparallel dikes. Low-dipping swarms near the contact are scarce.

Thirteen late-emplaced bodies were mapped in the eastern half of the batholith. The largest body is two-thirds of a mile long and half a mile wide and in plan view is shaped like a stubby lens. Three are of about the same shape but are somewhat smaller, and the remainder are thick dikes as much as a mile long. Where planar alignment of mafic minerals can be detected, it may be either parallel or oblique to the contacts of the bodies, but it always parallels the flow structure in the adjacent part of the batholith. This is particularly striking in two large dikes west and south of Feather Falls waterfall. Here the flow structures of the batholith apparently pass right through the dikes at right angles to their walls. At the ends of two of the lens-shaped bodies, the coarse, subparallel biotite flakes of the batholith pass for about 20 feet into the fine-grained aplite without changing their original alignment. In general, however, the contacts of these bodies are sharp.

Judging from their strongly preferred align-

ment at right angles to the curving flow structures, forming a subradial pattern, the late-emplaced bodies in both halves of the batholith were controlled by radial fractures resulting from upward pressure of an underlying mobile core. The few dikes in the east half of the batholith further the idea that the east half solidified early, thus being isolated to a large degree from the source of pegmatitic fluids. As shown by the orientation of their planar structures, the larger bodies may be of recrystallization or replacement origin, but the sharply bounded dikes of the west half bear no suggestion of such an origin.

Slip structures.—Scattered in the outer part of the batholith are fairly continuous, dark schistose layers of crushed granitic rock from 1 mm to several inches thick. They are either parallel to the planar-flow structures or cut them at low angles, and they distinctly offset the aplite and pegmatite dikes. Measured offsets on several thin smears are about a foot, but the thicker layers apparently represent far greater movements. Wherever the relative offset could be detected, the upraised block or slice was the one nearer the center of the batholith. As already noted, these slip surfaces account for some of the acutely angular junctions between flow layers.

Where late slippage is more evenly distributed in the rocks, it has produced a protoclastic gneissose structure and streaking of

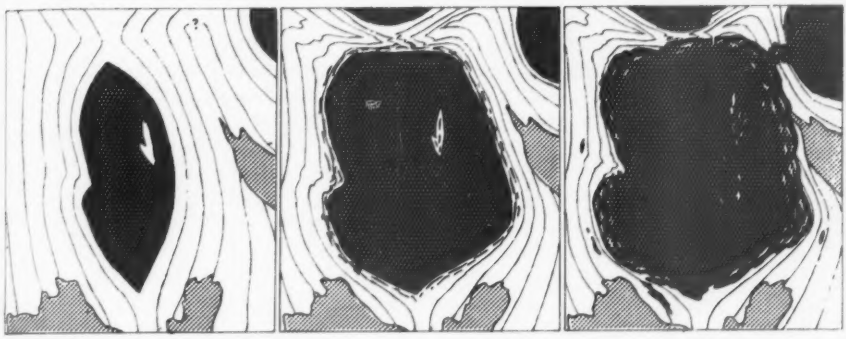


FIGURE 7.—GROWTH OF BALD ROCK BATHOLITH AND ADJACENT PLUTONS

Left: initial concordant intrusion of magma, with Swedes Flat pluton crystalline. Middle: approximate limit of concordant intrusion. Right: limit of discordant growth by stoping and assimilation. Darkest pattern = magma; lighter stippling = solid plutons; lines = strike of bedding and foliation.

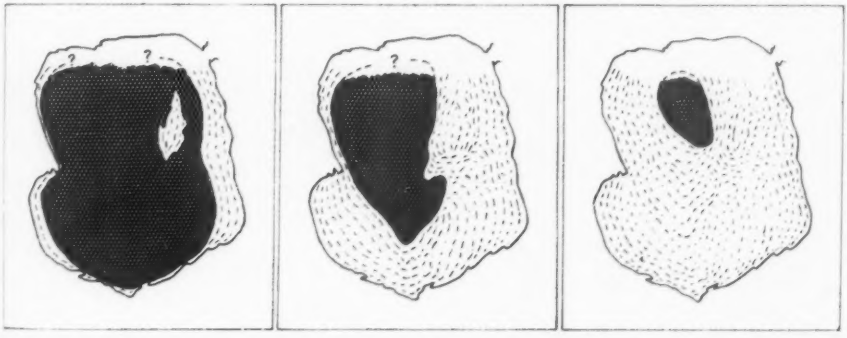


FIGURE 8.—CRYSTALLIZATION OF BALD ROCK BATHOLITH AS SHOWN BY FLOW STRUCTURES

Left: relation resulting from early rapid growth on east and west borders. Middle: trondhjemite magma, showing probable intrusive relations to southeast. Right: core of leucotrondhjemite magma. Dark pattern = magma; dashes = strike of flow surfaces.

biotite. These structures are especially clear and quite widespread near the west border of the intrusion between Lake Madrone and Canyon Creek; elsewhere they occur sporadically within

trondhjemite, and granodiorite, tonalite. In all of these the ratio of plagioclase to potash feldspar is greater than 2:1. The leucotrondhjemite consists of quartz and sodic to inter

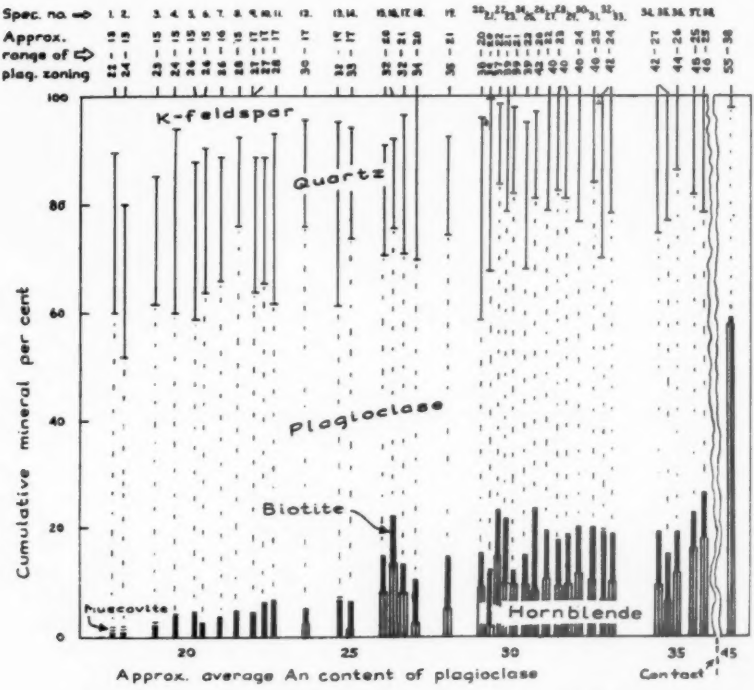


FIGURE 9.—MODES OF THIRTY-EIGHT BALD ROCK BATHOLITH ROCKS

Typical amphibolite to right of batholith contact. Specimen numbers shown on Plate 1 and Figure 10. 1. = Hietanen's analysed leucotrondhjemite; 11. = Turner's analysed trondhjemite.

1 mile of the contact and near the sharp keel in the batholith flow structures. The plunge of the protoclastic lineation, consisting mainly of smeared-out biotite grains, is down the dip of the planar flow structures, except near Canyon Creek where it is as low as 18° and is directed along and under the large projection of crumpled country rock and migmatite that occurs there. This lineation parallels the lineation of crumpling or of amphibole prisms in the near-by country rock.

Petrography

General.—Bald Rock batholith is made up of four kinds of granitic rocks: leucotrondhjemite, mediate oligoclase with subordinate microcline and less than 5 per cent biotite. The trondhjemite differs only in having more than 5 per cent biotite and slightly more calcic plagioclase. The average granodiorite consists essentially of quartz and calcic oligoclase or sodic andesine with less than 10 per cent microcline and about 8 per cent each of hornblende and biotite. The average tonalite contains only a trace of microcline and consists of quartz and either calcic oligoclase or sodic andesine with about 10 per cent each of hornblende and biotite.

Gradational nature of rocks.—The modes of these rocks show them to be essentially gradational, though sporadically so (Fig. 9). The modes were computed by the Rosiwal metho

with a determined error of about 5 per cent. They were used to key the mapping of the rock types of the batholith. Obviously, the number of modes measured is small compared to the size of the intrusion; however, the mafic content of the rocks gives a fairly reliable picture of their entire mineral content, and thus the granitic rocks can be mapped approximately by carrying typed samples in the field. The gradational contacts could not be mapped as lines; the color patterns of Plate 1 indicate the approximate boundaries between rock types. The trondhjemite-leucotrondhjemite boundary is accurate to within 1000–2000 feet. Small areas of trondhjemite probably occur within the leucotrondhjemite near its west border. The east and south boundary of trondhjemite and granodiorite are set within about 1000 feet, and locally much less; the west contact of these rocks, which is more broadly gradational, is accurate to only about 2000 feet. The least reliable relation is that of tonalite and granodiorite in the eastern half of the intrusion since few modes were measured here, and granodiorite is locally as mafic as tonalite. Undoubtedly, a few more patches of tonalite might occur in the granodiorite away from the border zone. The width of the border zone of tonalite on the south and west sides is reliable, in most places, to within 1000 feet. The distinct thinning of the tonalite border north of Feather Falls village on the east side, and north of Canyon Creek on the west side of the batholith, is well established. The gradational nature of almost all the rock boundaries is also considered well established. This was checked, after completion of the mapping, by several traverses across well-exposed parts of the batholith, and substantiated by the modal analyses. The rocks are so broadly gradational in most parts of the intrusion that its complex nature was not suspected until about half of it was mapped structurally. It was not until the writer saw Hietanen's chemical analysis of the leucotrondhjemite (Fig. 11) that the intrusion's great compositional range became apparent.

Megascopic characteristics.—All the rocks are medium- to coarse-grained and, except for those affected by protoclasis, all are hypidiomorphic granular. All the rocks show at least a crude planar flow alignment of hornblende, biotite, and plagioclase; though this is not obvious in every hand specimen, it is apparent in outcrops. The protoclastic rocks are distinctly foliate and show such clear signs of deformation as crude augen structures and linear biotite smears. Generally, the leucotrondhjemite and trondhjemite are uniform in appearance wherever examined. The tonalites, and to a lesser degree the granodiorites, show local variation in the amounts, kinds, and shapes of their dark mineral grains. Near the south end of the keel-like mass, for instance, tonalite with large scattered biotite flakes and subordinate large hornblende prisms grades in about 50 feet to tonalite with the same large biotite flakes, but with clusters of minute hornblende grains, and this in turn to tonalite in which both minerals are of medium size and equal amount. In several places in the keel and in the border zone, a distinct compositional banding in the tonalite lies athwart the flow structures and probably represents relic layering of large, half-digested inclusions of migmatite. In some places, there are sharp contacts between slightly different textural types of tonalite, but these contacts, when traced laterally, disappear. In a few places distinctive tonalites form ghost-like breccias, which look like swarms of nearly faded inclusions. The common areal distribution of inclusions and heterogeneous tonalite suggests the tonalite has been formed by varying contamination of trondhjemite by the country rock. If such is the case, most of the included material has been thoroughly digested or else made over into igneous-looking rock. The microscopic nature of the granitic rocks adds strongly to this point.

Microscopic characteristics.—The microtextures of all but the protoclastic rocks are distinctly hypidiomorphic granular. Plagioclase has strikingly good crystal form in most sections, commonly with well-developed Carlsbad habit. Combined progressive and oscillatory zoning of plagioclase is common to nearly all the batholith rocks. Most oscillatory rings are neatly concentric to the crystal form and represent small gradational decreases and then sharper and somewhat smaller increases of the anorthite content. Here and there, a ring cuts sharply across earlier rings. Rarely, veins

of oligoclase cut into a crystal core from an oligoclase rim, but albite does not occur in this relation. As many as 35 rings are present, though the average is 5–8. It was not feasible to survey accurately the relations between zoning and rock composition, but enough work was done to allow several general conclusions. (1) The grains with the most rings occur in the tonalites, but there is no complete correlation between the number of rings and rock composition. (2) Different grains within any rock sample show different numbers of rings, and the range differs greatly from rock to rock without apparent regard to rock type. (3) In a few rocks there are many grains that show only smooth progressive zoning. In several of the most deformed protoclastic rocks from the west border zone, plagioclase is unzoned, presumably because these rocks have almost entirely recrystallized. Albite twins are thin and rather sparse in the leucotrondhjemite and much of the trondhjemite, whereas they are common, quite broad, and joined by at least Carlsbad and pericline twins in the granodiorite and tonalite.

Quartz and potash feldspar, chiefly microcline, are distinctly interstitial. In the rocks in which they are most abundant, they each form poikilitic grains or large aggregates that have clearly replaced parts of the other minerals. Quartz-oligoclase myrmekite is common at contacts between plagioclase and microcline. Microcline is rarely and sparsely microperthitic. Subhedral dark-brown biotite is present in all the rocks. Hornblende, green to gray green in *Z*, is subhedral in the tonalites and most of the granodiorites but is shredded and scarce in the trondhjemites and leucotrondhjemites. Hornblende is rarely rimmed by biotite. Sphene, allanite, apatite, magnetite, and ilmenite are widespread accessories and are much more concentrated in some specimens than others. In several cases they either formed or recrystallized late in the growth of the plagioclase grains, for they are distinctly molded along plagioclase cleavage planes. Coarse muscovite occurs here and there in the leucotrondhjemite. Chlorite, leucoxene, clinozoisite-epidote, sericite, and albite are very sparse though widespread products of what must have been a mild deuteric alteration.

Composition.—A chemical analysis of leucotrondhjemite from Little Bald Rock was given by Hietanen (1951, Table 1, analysis 19), and one of trondhjemite from a point 1000 feet southwest of Enterprise, by H. W. Turner (1894, p. 482, analysis 95). These analyses appear on the variation diagram of Figure 11, and the modes of the same rocks are noted in Figure 9. The compositions of the batholith rocks are generally, though erratically, gradational. The change in anorthite content of plagioclase is apparently more consistent and gradational than the overall mineral per cents, and therefore it was used as the abscissa of the modal graph. It was not possible to measure the true average composition of the plagioclase, or even the composition of the smallest cores and most sodic rims of most grains. Therefore, measurements were made on the core that represented about the inner fifth (by volume) of the crystal and on the rim that represented about the outer fifth. These values are taken as the approximate range of zoning, while their average is used as an approximate average composition of the grains. Extinction angles were measured without the universal stage. The only check on the reliability of these figures is afforded by the normative plagioclase composition of the two analyzed specimens; these checked within 3 per cent An of the measured values. The writer suggests that the error for any one of the other specimens is about twice as great. In Figure 10 the variation of anorthite content of the plagioclase is plotted against the relative distance of the sample from the rock boundaries. Here the graph widths of each rock band are proportional to the approximate average width of the actual zones of the batholith. The general relations are further evidence for the sporadically gradational nature of the batholith rocks. The spread of plagioclase compositions in the tonalite band correlates well with the megascopic textural heterogeneity of these rocks. This spread is small compared with the great jump to the calcic plagioclases of the contact amphibolites.

Petrography of dike rocks.—Most of the dikes of the contact migmatites are tonalite and mafic granodiorite. These rocks are like those of the batholith except that they are

somewhat finer-grained. Many of the granular dikes have a heterogeneous distribution of mafic minerals; some grade to leucocratic stringers and blebs. Here and there the dikes are more basic than any of the batholith tonalites; the most basic rock sectioned is a quartz-bearing andesine-hornblende diorite.

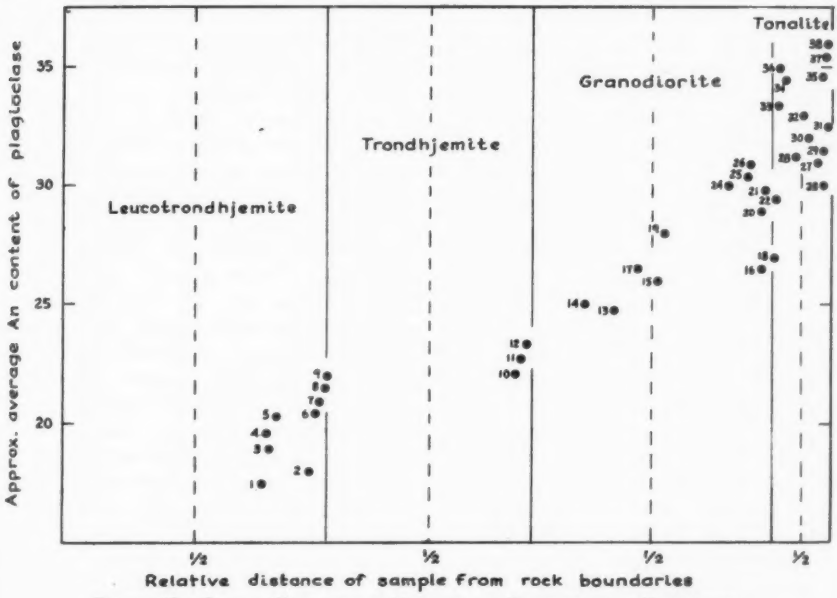


FIGURE 10.—SPATIAL VARIATION OF PLAGIOCLASE COMPOSITION IN BATHOLITH
Specimen numbers appear on Plate 1 and Figure 9.

The aplite and pegmatite associated with the contact migmatites are essentially like those of the dike swarms. These rocks average about 30 per cent quartz, 40 per cent microcline, 30 per cent albite (An_{5-10}), and small amounts of muscovite, biotite, and pink garnet. The aplites are allotriomorphic granular, while the pegmatites are generally hypidiomorphic granular.

The lens-shaped bodies and the thick dikes in the west half of the batholith consist mainly of fine-grained granodiorite and tonalite. Mineralogically, they are about the same as the corresponding types of the batholith. These rocks locally grade to aplite which contains less potash feldspar than the aplite of the dike swarms. The scarce, dark dike rocks are fine-grained hornblende-biotite tonalites and hornblende diorites. Most of these are characterized by zoned, euhedral hornblende phenocrysts and strongly zoned subhedral plagioclase, but some are allotriomorphic granular, and closely resemble the amphibolites. Probably specimens could be collected from the various late dikes and other masses that would show every gradation from aplite through fine-grained granodiorite, tonalite, and diorite to amphibolite. Volumetrically, the basic representatives of this dike sequence make up a minute part of the total.

Emplacement of Bald Rock Batholith

Interpretation of structures.—The structures of Bald Rock batholith clearly indicate that it was emplaced by forceful injection of a mobile magmatic body. The principal lines of evidence are: (1) the strikingly abrupt manner in which the country rocks have been pressed aside; (2) the nearly continuous occurrence of injection dikes and breccias in the adjoining wall rocks; (3) the presence of moved inclusions within the batholith; (4) the nearly universal

occurrence of planar-flow structures that show an integral, though curious, pattern within the batholith, and (5) a pattern of dike swarms and slip structures consistent with a late upward bulging of the batholith's core. The outward bulging of the country rocks accounts for about three quarters of the mass at the level of exposure, but the remaining quarter must have been emplaced in some other way. Since only inclusions and small parts of the wall rock were granitized, and since the contact migmatites and outliers consistently suggest dilative movement of slabs of country rock toward the batholith, the best possibility is piecemeal stoping. The local slip structures in the batholith and in the contact migmatites certainly indicate that the batholith tended to rise intermittently when its border was nearly crystalline; however, so much of the flow structures, inclusions, and migmatites are undeformed it seems likely the great lateral forcing of the magma ceased before the eastward and westward discordant extensions were completed. The writer believes the flow structures and their overall pattern can be explained only by assuming that the magma was mobile during the growth of the batholith and that grain orientation took place when a zone of low mobility slowly grew into the intrusion from its walls.

Interpretation of compositional variations.—Both structural and petrographic features indicate that the compositional variations of the batholith have been produced by contamination of trondhjemite magma by the stoped metabasaltic country rock. Perhaps the most important feature is the overall concentric arrangement of essentially continuous zones of granitic rock which become progressively more basic toward the contact. The generally gradational nature of the junctions between these rocks further supports the possibility of contamination. Of great importance in this regard is the thickness of the tonalite-granodiorite rim relative to the amounts of missing country rock. By far the greatest thickness of basic rocks occurs in the east half of the intrusion where the thickest crescent of metamorphic rocks has disappeared and where the flow structures indicate that a roof pendant has added to the amount of contamination. The next greatest crescent of tonalite and granodiorite lies in the southwest sector of the batholith where a sizable slice of country rock has been removed. It is striking that the tonalite rim is very thin in the two places where quartzofeldspathic metamorphic rocks are crosscut by the batholith (Fig. 10, specimens 16 and 18). The most obvious case is along the northwest contact between Lake Madrone and Canyon Creek, where metadacite and metarhyolite, in part migmatized, strike into the only consistently leucocratic border zone of the batholith. Again, just east of Feather Falls village, granodiorite almost impinges the contact where a band of metasandstones and metasiltstones is cut off by the intrusion. However, the biotite content of the granitic rocks does not increase at this place, and there are minor features elsewhere that indicate that the correspondence between compositions of contact rocks and of near-by granitic rock must be considered at best a general thing.

Contamination is further indicated by the heterogeneity of the tonalite and the granodiorite. Inclusion swarms, restricted to the tonalite border and the keel, show varying degrees of conversion to tonalite. Hazy bodies of more mafic or finer-grained tonalite that locally have sharp boundaries suggest a more advanced product of reaction between magma and inclusions. Large tonalite masses showing a distinct compositional banding indicate that sizable chunks of migmatite were locally stoped and made over without loss of continuity. The probable beginning of such a process is seen 1 mile west of Enterprise, where a large mass of contact migmatite was almost stoped into the intrusion before the border zone froze.

Are the amounts and compositions of the several granitic rocks consistent with a hypothesis of contamination of trondhjemite by basic metavolcanic rocks? The problem is essentially one of determining the average composition of apparently contaminated rocks and comparing it with the chemical analyses of the trondhjemite and leucotrondhjemite and the approximate composition of the contact aureole. The generally steep dips of the batholith contact, through an exposed vertical range of 3000 feet, and the consistently steep flow structures indicate that the unit being tested does not

change rapidly in a vertical direction. On this basis, the result of graphical measurement of the volume per cent composition of the rock types concerned. The biotite and hornblende compositions were obtained by averaging several chemical analyses given by Nockolds

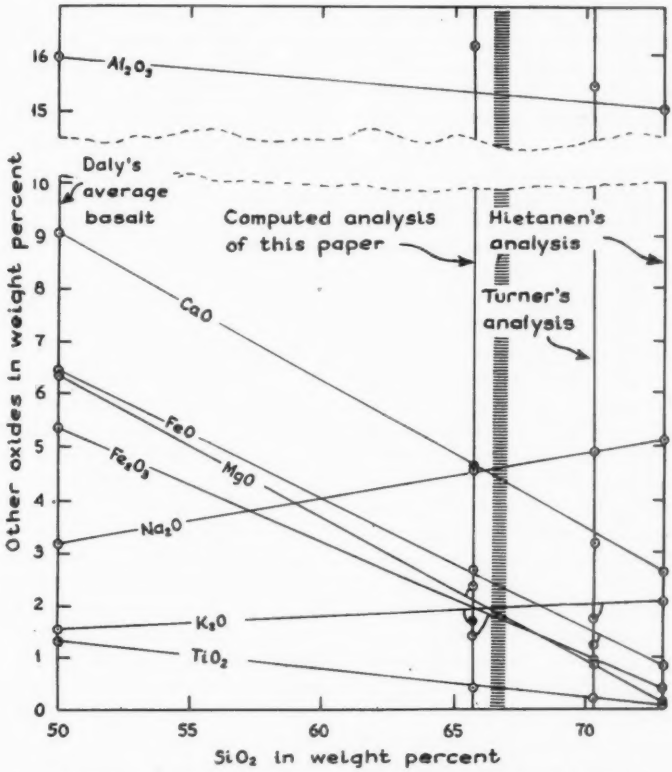


FIGURE 11.—VARIATION DIAGRAM OF BALD ROCK BATHOLITH ROCKS

Daly's average basalt recalculated to constituents used. Dashed band shows approximate ratio of mixing indicated by structural relations around batholith. Sources of analyses: Daly, R. A. (1914, p. 27); Turner, H. W. (1894, p. 482, analysis 95); Hietanen (1951, table 1, analysis 19).

batholith at and near the present level of exposure is: 20 per cent leucotrondhjemite, 22 per cent trondhjemite, 40 per cent granodiorite, and 18 per cent tonalite. By checking the modal analyses against the distribution of rocks, an approximate median mode was computed for the granodiorite and tonalite—19.6 per cent quartz, 58 per cent plagioclase (An_{27}), 3.3 per cent microcline, 7.6 per cent biotite, 10.8 per cent hornblende, and 0.7 per cent ore. A partial chemical composition was computed from this median mode by using mineral compositions that best fit the and Mitchell (1948, p. 560, 562) for minerals from similar rock assemblages. The plagioclase was allowed 0.8 per cent potash, and the microcline 0.9 per cent soda. Though epidote was originally calculated as plagioclase in the modes, it was considered to contribute 0.1 per cent ferric iron oxide to the analysis. The resulting analysis is shown on the usual silica variation diagram of Figure 11, where it is compared with Hietanen's and Turner's analyses of the trondhjemites and Daly's average basalt. The three batholith analyses produce remarkably straight-line trends. Only the alumina and

ferric iron oxide points fall far from the straight lines from leucotrondhjemite to average basalt. After trying various substitutions in this computation, the writer believes its only serious weakness is the difficulty of choosing an average composition of the country rocks. Judging from a general petrographic study, they were originally basalts or basic andesites; however, the latitude in composition here is admittedly large.

The ratio of mixture of the basalt and leucotrondhjemite end members (Fig. 11) is close to the ratio of mixing indicated by the structural relations of the batholith to its country rocks. If a small amount of country rock were added from above the exposed level, the ratio would be even closer.

Problems concerning contamination.—If contamination by basic metamorphic rocks is the most plausible origin of the diverse rock types of Bald Rock batholith, its mechanism must be deduced. The problem is dominantly one of reconstitution of plagioclase, which forms over half the granitic rocks. Plagioclase shows both progressive and oscillatory zoning in all parts of the batholith; its average composition changes gradually and considerably from the core of the batholith to the contact. Immediately across the contact, almost all the plagioclase of the amphibolites shows only progressive zoning and is far more calcic than that of the adjoining granitic rocks. Apparently the plagioclase grains of amphibolite inclusions have not been made simply more and more sodic before and during imbibition, but have been reconstituted into new oscillatory-ringed grains. Notably, the cores of these grains are rarely as calcic as those in the original amphibolite. The mafic minerals show the same sort of relations. Biotite is scarce in the amphibolite at the contact but is a common and coarse-grained constituent of the adjoining tonalite. Hornblende in the tonalite is far more coarse-grained and idiomorphic than in the amphibolite. Quartz is abundant out to the wall of the intrusion, or the contacts of the injection migmatites, but is scarce in amphibolite even a few inches from the contact. Obviously, whatever the mechanism of mixing, it required great mobility of the imbibed materials during their assimilation.

The possibility that the magma completely or largely melted the wall rocks and inclusions is put aside on theoretical grounds and because the vague heterogeneous bodies in the tonalite indicate that many inclusions were made over into tonalite without disintegrating. Bowen (1928, p. 175–223) emphasized that, when grains high in either a continuous or discontinuous reaction series are added to an acidic melt, an exothermic reaction will ensue whereby the grains are made over into the phases in equilibrium with the melt. Such reactions presumably took place here. Nockolds (1933) pointed out that Bowen's mechanism affords the energy for making over basic inclusions but does not allow contamination of the magma itself, and he suggested that the magma can be contaminated by a reciprocal reaction or exchange of materials between magma and inclusions, which may be quantitatively important if the exchanging materials are sufficiently mobile. He deduced that the exchanging medium within the inclusion was a volatile-rich solution. It appears that such a granitization of inclusions has taken place in Bald Rock batholith; however, either the magma was originally heated above the saturation point, or the reciprocal reactions were considerably exothermic. Otherwise the magma should have crystallized soon after it began being basified by its inclusions. The time relation seems especially critical. The intrusion shows, within the contact zone, the apparent anomaly of being unable to granitize much of its walls, but able to make over most of the included wall rock to tonalite and to mix the products quite well. As already noted, stoping and granitization at the contact must have stopped when the contact zone froze, but the remaining magma was reactive long enough to make over most of its inclusions. The occasional dark and fine-grained inclusions in the tonalite are difficult to explain; perhaps they are wall-rock fragments whose textures, mineral compositions, or surface characters opposed reaction with the magma. Perhaps they were moved into the magma just before it solidified.

The way in which the flow layers locally cut across the gradational rock boundaries poses a considerable problem. If the writer's interpretation of the flow structures is correct, this must mean that immobile tonalite was forming near the contact while granodiorite

was forming somewhat farther from the contact, and, in some cases, while trondhjemite was forming still farther from the contact. Either the interior of the batholith was cooler than its rim, or trondhjemite, with plagioclase zoned out to An_{17}, crystallized at the same or a higher temperature than tonalite with plagioclase zoned out to about An_{24}. If the batholith is an integral unit of magmatic growth, as its structure and petrography indicate, the writer feels there was a positive temperature gradient between the core and the contaminated border, caused by the increasingly exothermic nature of the reactions toward the basic rim of the intrusion. Though the temperature rose, the amount of reacting magma dwindled. Heat loss to the aureole finally stopped the corrosive activity of the intrusion. Such a delicate balance between exothermic reaction, amount of contamination, and heat loss to the aureole explains the local abundance of oscillatory zoning in the tonalites—the oscillations record the local gain and loss of energy and of material by the dwindling magma.

The nature of the late-formed flow structure indicates that even in the most contaminated parts of the border zone most of the tonalite was once mobile. Here, the original ratio of trondhjemite magma to included solid material would have to be about 1:1. If the magma were saturated before intrusion, it should have begun crystallizing rapidly after beginning to mix with such a load of basic rock. Thereby, the border zone should be far more heterogeneous than it is. The time relation in this regard has already been emphasized. The prolongation of time necessary for thorough reaction and mixing in a mobile state requires either that the original magma was superheated or that there was some outside source of energy. The question of superheat is left as problematical. Possibly hot volatiles, streaming through the intrusion from below, helped prolong its period of reaction and consolidation.

Contact-Metamorphic Rocks

General Statement

Contact-metamorphic rocks underlie about 30 square miles of the area mapped, chiefly as a continuous band around Bald Rock batholith. On the west and southeast sides of the batholith, where the contact rocks can be considered results of its heating effects, the aureole is 1–2 miles wide. On the south and east sides of the batholith, Swedes Flat pluton and the small pluton underlying the Swains Hill-Big Creek tract undoubtedly have contributed to the contact metamorphism of the metavolcanic rocks.

The metavolcanic contact rocks were mapped in two zones; the outer zone consists chiefly of dark-gray or black fine-grained hornfels, the inner of coarse hornfels and amphibolite locally migmatized by tonalite much as are the mixed rocks that border Swedes Flat pluton. The zones can be mapped only across metavolcanic rocks, for the metasedimentary rocks are essentially schistose throughout the contact aureole. The contact between the inner and outer zones is placed where metamorphic amphibole can be seen megascopically; it is gradational through many tens of feet of intermixed fine- and medium-grained layers. The contact between the outer zone and the low-grade regional terrane is also gradational but in most places can be set quite closely. This contact is particularly well marked by the topography, for the tough hornfels stands as a high ridge above the more subdued terrane of the low-grade, regionally metamorphosed rocks.

Structures

Bedding.—Bedding is discernible in most parts of the outer zone. The beds are of the same order of thickness as those of the regionally metamorphosed rocks, and though they are not nearly so obvious as in the low-grade terrane they show up on large, slightly weathered outcrops. Bedding passes with conformity from the regional terrane into the outer hornfels zone. In the inner zone, a large-scale compositional layering, assumed to be largely bedding, is conformable with bedding of the outer zone. In some places the layers are offset along shears that cut them at low angles, and in other places there are clear-cut boudinage relations between brittle quartz-rich hornfels and less competent amphibolite. Apparently relic bedding in the inner zone has been locally modified by slip movements that are about parallel to it, and thus some of the compositional layering is of metamorphic origin.

Foliation and lineation.—In the field, foliation in fresh metavolcanic hornfels of the outer zone can be detected by the tendency of some beds to split with a hackly fracture that consistently parallels the bedding. Weathered hornfels exhibits a crude fissility. As can be seen under the microscope, this fissility depends on grain orientation that is mimetic to schistosity of the low-grade terrane. The inheritance of regional schistosity in contact metamorphism is shown more obviously by the metasedimentary rocks in the outer zone, for these are consistently schistose and only somewhat coarser-grained than their regional equivalents. In many places in the outer zone the quartz-mica schists show a steeply plunging lineation of crinkle axes, intersections of slip cleavage with bedding, and streaks of white and dark mica. In the metavolcanic hornfels, the only lineation is a vague parallelism of minute hornblende prisms. There are no clear-cut folds in the outer zone to which to relate the lineations.

The metavolcanic rocks of the inner zone show a clear and widespread lineation but only a crude schistosity defined by hornblende prisms or, less commonly, by foliate aggregates of granular hornblende. Schistosity parallels the gross layering of the rocks. Simple, foliate gneissose structure is rare; wherever gneisses occur, they are crumpled. Alignment of hornblende prisms, always in the plane of foliation, is the most common linear structure, and here and there it parallels streaking of hornblende and plagioclase on foliation planes, and crumple axes of migmatites. The metasedimentary rocks in the inner zone are schistose and, near the batholith contact, commonly gneissose. In conglomeratic beds, the axial ratios of stretched pebbles range from about 1:2:5 to 1:3:10, paralleling both the foliation and lineation of near-by metavolcanic rocks. Lineations in the quartz-mica schists consist of mica streaks on foliation planes, crumple axes in coarse micaceous layers, and intersections of slip cleavage and schistosity. These lineations are essentially parallel to each other as well as to the hornblende prisms in the metavolcanic rocks.

Bedding and schistosity strike about parallel to the contact of Bald Rock batholith and dip steeply away from that body in most places. The widespread lineations of the inner zone show a strikingly consistent pattern, plunging steeply away from the batholith in all parts of the aureole except for the tract of strongly crumpled rocks along the west-central border of the batholith. Here, the crumple axes nearest the contact plunge at low angles along the trend of the contact. Away from the batholith, they swing gradually to coincide with the steep outward plunge shown by the rest of the aureole. These crumpled rocks lie somewhat over the adjacent part of the batholith. They were probably crumpled by pressure from the core of the batholith against its nearly crystalline border.

Randomly distributed migmatites.—Veins and dikes of tonalite of recrystallization and replacement origin are widespread through the inner zone of the aureole. On Plate 1, they have the same color as the contact migmatites, but in most places they can be distinguished by their structural pattern. They are mapped only where they are well exposed or where they could be inferred from scattered outcrops. If unexposed areas could be seen, their frequency would probably be doubled. Near the contact, they cannot be distinguished everywhere from contact migmatites; however, they are not concentrated at or near the contact. They are most abundant in the broad tract of high-grade rocks south of the batholith, lying as far as 2½ miles from the contact.

Structurally, most of these migmatites are identical to the diked and veined terrane that borders Swedes Flat pluton, though nowhere do they form large continuous masses of granitic rock. The granitic rocks occur as weblike veins, irregular dikes, and inclusion-charged nodes (Pls. 4, 5). They are rarely foliate, crumpled, or otherwise lineate. Sharp contacts are common, but hazy gradations to porphyroblastic amphibolite occur. In a few instances small inclusions have been moved in the thin dikes. Under the microscope these dikes show a flow structure parallel to the dike walls, contrasting with the random fabric of the greater number of dikes (Pl. 5, fig. 2). Apparently the randomly distributed migmatites were in small part magmatic.

Late phyllonite sheets.—Thin sheets of dark

phyllonite cut the contact aureole in many places, appearing as nearly vertical black bands in all the contact rocks, including the heterogeneous diorites and gabbros. They are generally many dozens of feet apart but locally

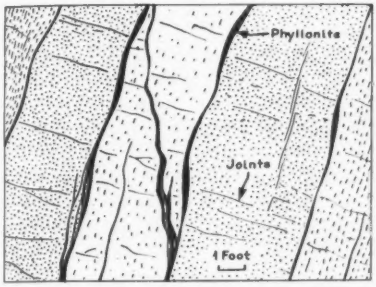


FIGURE 12.—PHYLLONITE SHEETS CUTTING LAYERED HORNFELS AND AMPHIBOLITE
The lined patterns indicate foliate amphibolite, the dotted patterns hornfels. Vertical exposure on South Fork of Feather River, northwest of Forbestown.

are only about a foot apart (Fig. 12). Thickness ranges from paper thin to about 3 inches and varies greatly within a single sheet. They are parallel to the bedding and foliation or cut them at angles of less than 30°, so that their aggregate pattern is essentially concentric to the batholith. Since they consist of crushed and only slightly recrystallized hornfels and amphibolite, and since they are concentric to the batholith, they are probably slip structures produced by a late shouldering of the intrusion, when both the intrusion and its aureole had cooled considerably.

Petrography of Contact Aureole

Textures of metabasaltic rocks.—In the outer zone, particularly near the regional-metamorphic terrane, the hornfels has microtextures relict to textures of the low-grade schists and massive metabasalt or metadolerite. Recrystallized greenschists retain considerable grain orientation, metadolerites are still recognizable by their coarser grain, and old phenocrysts in metadacite show up clearly. In all the rocks, the dusty, fuzzy groundmass has changed to a cleaner, more evenly granoblastic one (Fig. 13 [middle]). This change is due chiefly to the aggregation of feathery actinolite into distinct, elongate hornblende prisms and of fine leucoxene into sphene and ore. In most parts of the outer zone, the hornblende prisms are about .05 mm by .3 mm; plagioclase and quartz are granular and about .01 to .1 mm in diameter. Hornblende is locally porphyroblastic in .5-mm prisms.

The contact between the outer and inner zones is gradational with regard to the microtextures of the rocks. Hornblende in the inner zone ranges from about .1 to .5 mm; and plagioclase from .1 to .2 mm. Grain sizes do not gradually increase toward the batholith contact but go up sharply wherever the amphibolites are segregated, or impregnated with tonalite veins. Besides the overall increase in grain size, the principal textural change from the outer zone rocks is that hornblende becomes more equigranular. Many of the high-grade amphibolites are coarse-grained granoblastic rocks (Fig. 13 [right]). Near granitic veins and dikes of the randomly distributed migmatites, amphibolite is likely to be segregated, and here hornblende is coarsest (up to 1 cm) and subhedral or, less commonly, sieved. As already noted, hornblende has a strongly preferred crystallographic orientation in the widespread lineate rocks of the inner zone. Under the microscope, {100} lies in or close to the plane of schistosity, while the prism axis defines the steep lineation. Even in rocks that show no lineation of grain shape, more than half the hornblende grains lie close to this preferred direction.

Mineralogy of metabasaltic rocks.—The mineral assemblages of rocks collected from all parts of the contact aureole show a general rise in metamorphic grade as the batholith contact, or areas of migmatites, are approached. Figure 14 shows the change in composition of plagioclase in rocks of basaltic composition, plotted against relative distance from the batholith and aureole contacts. The most calcic feldspar occurs in and near the randomly distributed migmatites. Here, plagioclase is notably more zoned than in the rest of the aureole. Hornblende makes up 40–60 per cent of the metabasaltic rocks. Hornblende is consistently blue green in Z and darker than the actinolite of the regional terrane. Its color-

ing is somewhat irregular in the outer zone but becomes more even as the batholith is approached, and its intermediate index of refraction increases from about 1.655 to 1.67 in the total range of the aureole. A dark amphibole, probably ferrohastingsite, is abundant in two places in the inner zone and in one inclusion. Diopsidic (?) augite is present in several specimens from near the contact. Epidote is a part of the equilibrium assemblage of most of the outer-zone metabasaltic rocks, though it is less abundant than in the regional terrane and contains less iron. It is rare near the boundary between the zones and occurs in the high-grade rocks only as veins or segregations that are probably hydrothermal. Chlorite is present in a few of the outer-zone metabasalts, but it is not possible to determine whether it is an equilibrium mineral or a metastable relic. Olive-green biotite appears stable in the outer zone. Sparse red-brown biotite in the amphibolite of the contact migmatites apparently was produced by a meager addition of potash from the batholith. Fine-grained red garnet occurs in some of the smeared amphibolite of the contact migmatites. The only common accessories of the aureole are apatite, opaque iron ores, and sphene, which coarsen toward the batholith.

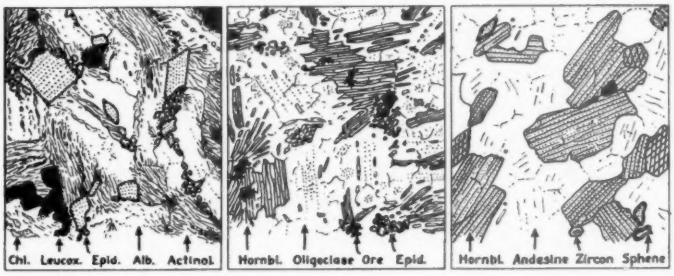


FIGURE 13.—TEXTURAL CHANGES IN MASSIVE METABASALTIC ROCKS OF REGIONAL TERRANE AND CONTACT AUREOLE

Left: metadolerite of regional terrane. Middle: probable metadolerite of outer contact zone. Right: amphibolite of inner contact zone.

Metadacites.—Besides metabasaltic rocks, the only volcanic rock type repeated throughout the aureole is a leucocratic flow or dike rock that was probably once a dacite. In the outer zone its plagioclase ranges from about An_6 to An_{18}, and in the inner zone it does not rise above An_{26}. Quartz makes up about 25 per cent of these rocks. Olive-green biotite, chlorite, and hornblende, similar to those of the metabasalts, are the mafic minerals in the outer zone, and hornblende and green-brown biotite are the mafic minerals of the inner zone. Epidote is scarce, even in the outer zone. Opaque iron ores and sphene are constant accessories.

Petrography of metasedimentary rocks.—The metasedimentary rocks are schistose in almost all parts of the contact aureole, except for one occurrence of granoblastic tactite and a few outcrops of black hornfels. Under the microscope, even the hornfels shows a high degree of grain orientation, presumably mimetic to an original phyllite schistosity. The grain size of biotite in the schists increases from about .03 mm in the phyllites of the regional terrane to about .2 mm in the outer part of the contact aureole and to as much as 1.5 mm near the batholith. Porphyroblastic hornblende occurs in mica-oligoclase-quartz schists that are probably derived from tuffaceous or limy sediments. The coarsest rocks, which always occur within a few hundred yards of the batholith, are commonly gneissose and migmatitic, though much diaphthorized.

In the quartz-rich, hornblende-poor schists, sodic oligoclase occurs in the outer zone, and intermediate oligoclase in the inner zone; in schists with much hornblende, intermediate oligoclase occurs in the outer zone, and calcic oligoclase or andesine in the inner. Except where abundant, hornblende is paler than in

the metavolcanic rocks. Biotite is pale red brown in the schists of the outer zone and medium to deep red brown in the inner zone. Garnet occurs locally in the inner zone schists.

grained quartz and feldspar, as though the dike rock had been protoclastically deformed. The other dikes and veins studied petrographically, representing 15 localities, show no sign

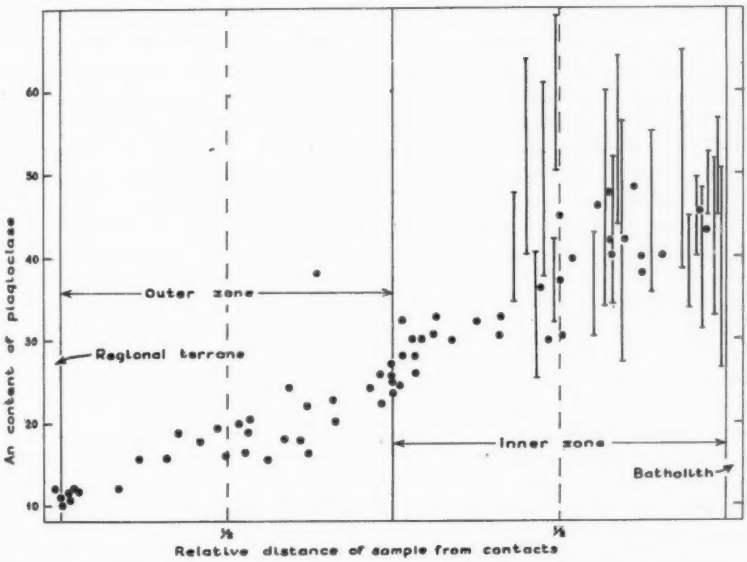


FIGURE 14.—DISTRIBUTION OF PLAGIOCLASE COMPOSITIONS IN METABASALTIC ROCKS OF CONTACT AUREOLE
Dots are approximate average compositions where zoning is slight or absent; lines show approximate range of progressive zoning.

Discrete masses of fine-grained white mica in the coarsest gneisses were probably once andalusite or some other high-grade aluminum-rich mineral; however, no relics of such minerals were found. Widespread diaphthoresis, presumably following migmatitization of the rocks, has largely reduced biotite and any other high-temperature aluminous silicates to muscovite, chlorite, and opaque ores. Local concentrations of tourmaline and apatite suggest at least some additive effects.

Petrography of randomly distributed migmatites.—The structures of the randomly distributed migmatites indicate that most of their tonalite veins and dikes are of replacement origin, but that a few are magmatic. These few, under the microscope, are hypidiomorphic granular, and their plagioclase tablets define a clear-cut flow structure parallel to the dike walls (Pl. 5, fig. 2). In two such dikes, the grains are bent, broken, and healed with fine-

of movement, though they are mineralogically identical to the ones with flow structure. Their textures range from hypidiomorphic granular to crystalloblastic. As in the Swedes Flat mixed rocks, quartz forms intergranular clots that have replaced parts of the plagioclase and hornblende; it forms 5–45 per cent of the rocks. Plagioclase is typically zoned in a progressive manner, and in three places it shows oscillatory zoning with as many as 8 or 10 rings. In the most calcic dike studied, the zoning is from about An_{65} to An_{45} and in other specimens it ranges down to about An_{40} to An_{25}. Hornblende is subhedral in most places and in all cases is exactly the same color as the hornblende of the surrounding amphibolite. In notable contrast to the contact migmatites, the veins contain neither biotite nor potash feldspar, and almost all the rims of their plagioclase grains are more calcic than those of the sheeted dikes of the contact migmatites.

The contact between hornfels or amphibolite and tonalitic vein rock is commonly sharp, though the country rock may suddenly become coarse-grained and somewhat segregated close to the veins. Locally, the amphibolite contains porphyroblasts of plagioclase (Pl. 5, fig. 3). The plagioclase in the amphibolite near tonalite veins is commonly progressively zoned and has about the same composition as that in the vein, in some cases being more calcic. Whereas hornblende makes up about half or somewhat more of the amphibolite, it forms only about 10 to 40 per cent of the dike rock; as in the Swedes Flat migmatites, it is greatly concentrated at some contacts. Apart from the relative amounts of the minerals, dikes and veins of tonalite have similar mineralogy to that of the surrounding amphibolite, in contrast to the abrupt difference between the batholithic tonalite and its bordering amphibolite.

Retrograde or hydrothermal effects.—The low-temperature mineral assemblage of the thin phyllonite layers indicates that they were formed by slip and recrystallization when the batholith and aureole were considerably cooled. Such low-temperature minerals as epidote, leucoxene, sericite, albite, chlorite, calcite, and zeolite also occur widely though sparsely through the aureole. They are associated chiefly with thin veins and irregular clots of quartz and epidote. Pyrite is common in the outer zone near Bidwell Bar. Large quartz veins occur in the inner zone near Canyon Creek and also near Forbestown, where they have been mined intermittently for their gold content.

Interpretation of Contact Aureole

Mineral-zoning and facies considerations.—There is clearly a progressive and somewhat sporadic change of mineral paragenesis between the regional terrane and the batholith or the randomly distributed migmatites. In rocks of basaltic composition the key minerals are plagioclase, epidote, hornblende, and perhaps chlorite. Chlorite is depleted near the outside of the outer zone, and epidote dwindles gradually as the anorthite content of plagioclase increases. Since epidote disappears at about the contact between the zones, the inner zone is essentially of amphibolite facies. Diopsidic (?) augite in several near-contact amphibolites andicates that pyroxene hornfels facies may have been reached here. The abrupt rise in inorthite content of plagioclase near the replacement migmatites may be due to higher temperatures there or, to concentration of Ca and Al from the recrystallizing tonalite veins.

The mineral assemblage oligoclase-epidote-hornblende in basaltic rocks of the outer zone does not fit the required albite-epidote-hornblende assemblages of Eskola's epidote-amphibolite facies, though it fits certain parts of that facies as described by Barth (1952, p. 339). Clearly the rocks mapped here lie between rocks of probable greenschist facies and amphibolite facies; the writer suggests that the temperature of their environment is the same as that required by Eskola's epidote-amphibolite facies, but that stress was low enough in the contact aureole to allow oligoclase to form instead of albite. Thus, the assemblage oligoclase-hornblende-epidote would be the contact or thermal equivalent of the assemblage albite-hornblende-epidote of dynamothermal metamorphism.

Mechanism of heat transfer to aureole.—The distribution of the contact zones indicates the zones have been formed by recrystallization of the low-grade regional terrane by heat from Bald Rock batholith and its neighboring plutons. The peripheral distribution of the Bald Rock batholith aureole gives the incorrect impression that it is much more limited than the intrusion. Actually, if the zones are projected through other than metabasaltic rocks, they cover about half as great an area as the batholith. If the stoped and assimilated rocks are added to these, the total area of contact-metamorphic rocks is about equal to that of the intrusive magma prior to contamination. The rise in grade within the aureole is sporadic, according to the anorthite content of rocks of approximately basaltic composition. It is suggested, therefore, that heat additions were largely from fluids traversing fractures and relatively permeable zones, rather than from simple thermal diffusion from the batholith walls. A fluid mechanism is particularly apparent in the randomly distributed migmatites. They show a sporadic distribution, clear-cut

fracture control, and a sharp increase in grain size, segregation, and apparent grade of the amphibolite around them. The peak of this contact metamorphism was apparently reached

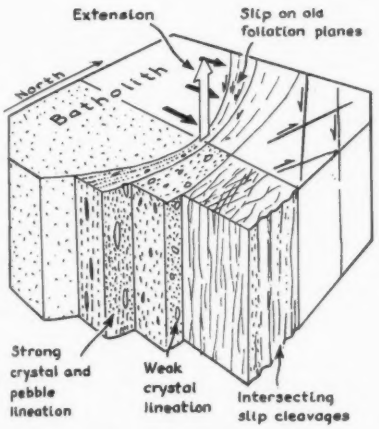


FIGURE 15.—RELATION OF LINEAR STRUCTURES TO BATHOLITH

Arrows show probable resolutions of intrusive force of Bald Rock Batholith. In front half of block, resulting linear structures are indicated.

where the fluids produced mobile tonalitic material in some of the replacement veins.

Structural effects of batholith on its country rocks.—The country rocks were forced strongly to the east and west during the initial intrusion of the batholith. Not only were their steep planar structures strongly bent, but the units nearest the east and west contacts were greatly thinned chiefly by a vertical extension. Can such an extension be the cause of the widespread lineations of the country rocks? The steep plunge of the lineations essentially parallels the axis of extension and the vertical axis of the batholith. Moreover, the degree of lineation increases toward the contact. The vertical long axes of stretched pebbles of conglomerates fit such a picture. The more common orientation of hornblende prisms parallels the stretched pebbles, but its mechanical significance can only be guessed. Presumably there could not have been enough rolling on a vertical axis to produce this strongly preferred prism alignment. It is perhaps no easier to picture its origin by plastic vertical extension, but some such streaming affect would be the only other kinetic basis for the lineation. The vertical crumple axes indicate local horizontal slip on the steep foliation planes. Perhaps the slip cleavage in the adjoining regional terrane is a further expression of the outward forcing of the batholith. In Figure 15 the linear structures are diagrammatically summarized in relation to the forces resulting from the intrusion.

The strong lineations and the thinning of the inner zone rocks emphasize one point in the development of the aureole; the randomly distributed migmatites were formed largely after the outward forcing of the contact rocks. These migmatites rarely show effects of either planar or linear distortion. The thin phyllonite layers that cut all the contact rocks must have been formed very late in the structural development of the aureole.

SUMMARY OF NEVADAN PLUTONISM

The broad structural relations between the regional terrane, the contact-metamorphic rocks, and both Swedes Flat pluton and Bald Rock batholith clearly indicate that regional dynamothermal metamorphism preceded igneous emplacement. Bald Rock batholith forced its own structural pattern both on the gross trends and the detailed fabrics of the surrounding metamorphic rocks. Undoubtedly this intrusion invaded a terrane of low-grade rocks; it can only be suggested that it was the nearly vertical foliation of the regional schists that allowed the intrusion to arrive rapidly, with little loss of heat energy, into a low-grade terrane. That it did arrive as a very energetic magma is indicated by nearly every aspect of its structure and composition, particularly by its high-grade contact effects, its apparent ability to have stoped and assimilated country rock equal to about a third of its volume, and the indications of a late-formed flow structure. Outside of rapid arrival, the only mechanism the writer can suggest that would help the intrusive magma retain a high temperature during ascent into a low-grade terrane is continuous heat transfer by volatiles streaming from below. Such volatiles may have aided greatly in the magma's imbibition of country rocks, and, where these high-temperature fluids

passed along fractures in the country rocks, they may have produced the labradorite-bearing tonalites of the randomly distributed migmatites.

The proximity of the several granitic plutons of the northwestern Sierra belt suggests a common origin. Apparently, the squarish shape of Bald Rock batholith was determined by the shapes and positions of the surrounding plutons (Fig. 7). The earlier emplacement of Swedes Flat pluton stopped its southern extension. Merrimac pluton and the large pluton to the east are structurally similar to Bald Rock batholith and probably coeval with it. At least Merrimac pluton has a trondhjemite core, though Hietanen (1951, p. 594) considers this a late separate intrusion. In any case the granitic rocks that she describes are, like those of Bald Rock batholith, relatively low in potash. Can two so different masses as Bald Rock batholith and the north end of Swedes Flat pluton be genetically related? The replacement migmatites of Bald Rock batholith may afford the answer, for they are structurally identical to the veined border of Swedes Flat pluton. Possibly Swedes Flat pluton was formed when an intrusion similar to Bald Rock batholith was stopped at a lower level, the replacement mass being formed by volatiles that rose from, through, and around the subjacent magma. Finally, are these several granitic bodies merely large cupolas of an extensive elongate pluton that underlies the northwest Sierra? The only direct evidence lies in the distribution of the contact aureoles: the clear restriction of high-grade rocks to the borders of the granitic masses indicates that if a regional batholith exists it must lie at considerable depth.

References Cited

Barth, T. F. W., 1952, Theoretical petrology: N. Y., John Wiley & Sons, 387 pages.

Becker, G. F., Lindgren, W., and Turner, H. W., 1895, Smartsville Folio: U. S. Geol. Survey Geol. Atlas, folio 18.

Bowen, N. L., 1928, The evolution of the igneous rocks: Princeton, N. J., Princeton Univ. Press, 332 pages.

Daly, R. A., 1914, Igneous rocks and their origin: N. Y., McGraw-Hill, 563 pages.

Eskola, P., 1925, On the petrology of eastern Fennoscandia, I, The mineral development of basic rocks in the Karelian formations: Fennia, 45, no 19, p. 1–93.

Ferguson, H. G., and Gannett, R. W., 1932, Gold quartz veins of the Alleghany District, California: U. S. Geol. Survey Prof. Paper 172, 139 pages.

Hietanen, A., 1951, Metamorphic and igneous rocks of the Merrimac area, Plumas National Forest, California: Geol. Soc. America Bull., v. 62, p. 565–608.

Hutton, C. O., 1940, Metamorphism in the Lake Wakatipu region: New Zealand Dept. Sci. and Indus. Res. Geol. Mem. no. 5, 90 pages.

Lindgren, W., 1900, Colfax Folio: U. S. Geol. Survey Geol. Atlas, folio 66.

Nockolds, S. R., 1933, Some theoretical aspects of contamination of acid magmas: Jour. Geology, v. 41, p. 561–589.

Nockolds, S. R., and Mitchell, R. L., 1948, The geochemistry of some Caledonian plutonic rocks: a study in the relationship between the major and trace elements of igneous rocks and their minerals: Roy. Soc. Edinburgh Trans., v. LXI, p. 533–575.

Saether, E., 1945, Studies on the igneous rock complex of the Oslo region; III, the south-eastern part of the Baerum-Sørkedal cauldron: Norske Videnks.-Akad. Oslo, mat.-nat. kl., Skr., no. 6, 60 pages.

Sugi, K., 1931, On the metamorphic facies of the Misaka series: Japanese Jour. Geology and Geography, v. 9, no. 1–2, p. 87–142.

Turner, F. J., 1933, The genesis of oligoclase in certain schists: Geol. Mag., v. lxx, p. 529–541.

—— 1935, Metamorphism of the Te Anau series in the region north-west of Lake Wakatipu: Roy. Soc. New Zealand Trans., v. 65, pt. 3, p. 329–349.

Turner, H. W., 1894, Rocks of the Sierra Nevada: U. S. Geol. Survey Ann. Rept. 14, pt. 2, p. 435–495.

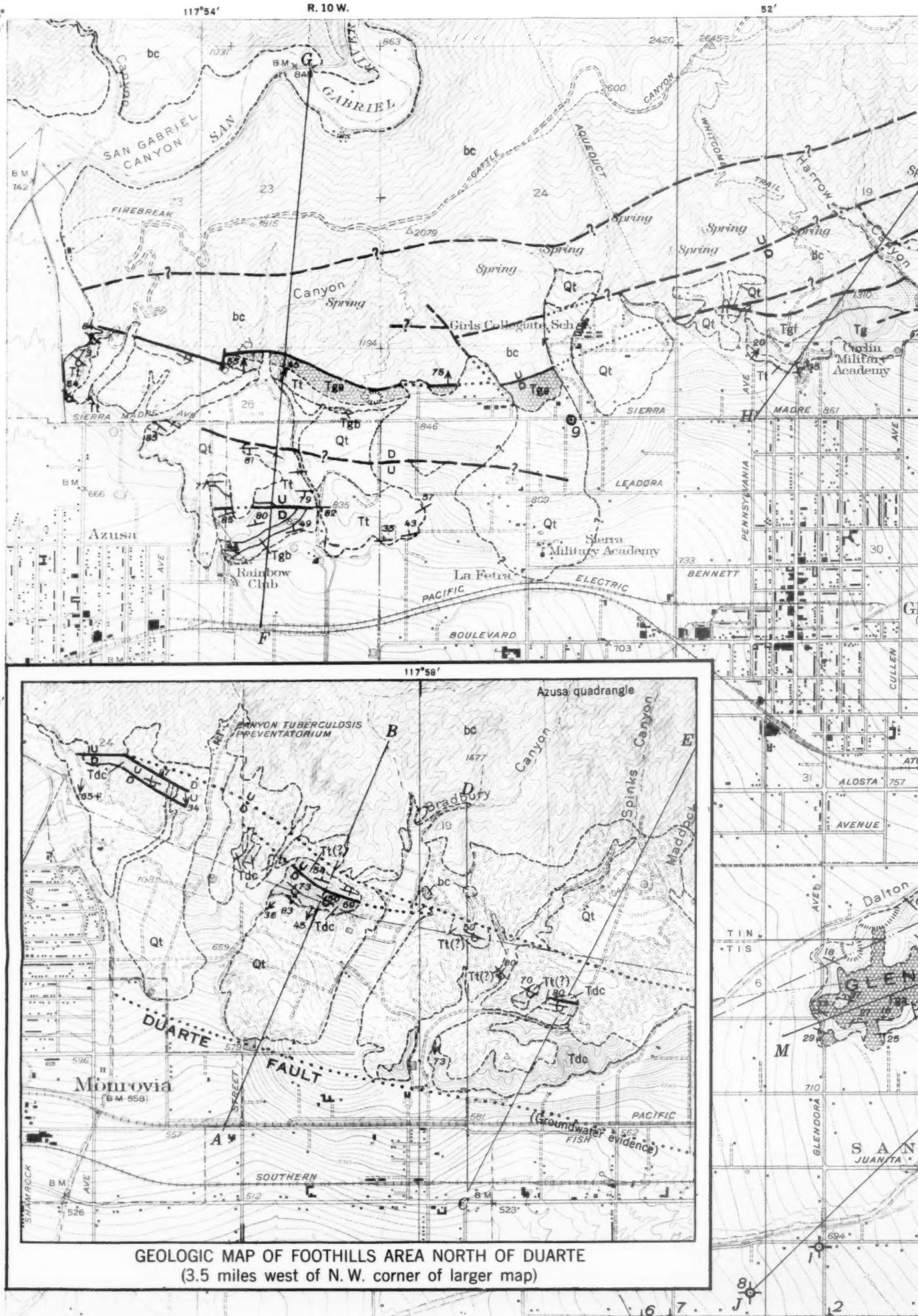
—— 1896, Further contributions to the geology of the Sierra Nevada: U. S. Geol. Survey, Ann. Rept. 17, p. 521–744.

—— 1898, The Bidwell Bar Folio (incl. Description of the Gold Belt, a supplement): U. S. Geol. Survey Geol. Atlas, folio 43, 4 pages.

Vogt, T., 1927, Sulitelmafeltets Geologi og Petrografi (with English summary): Norges Geol. Undersök., no. 121, p. 1–560.

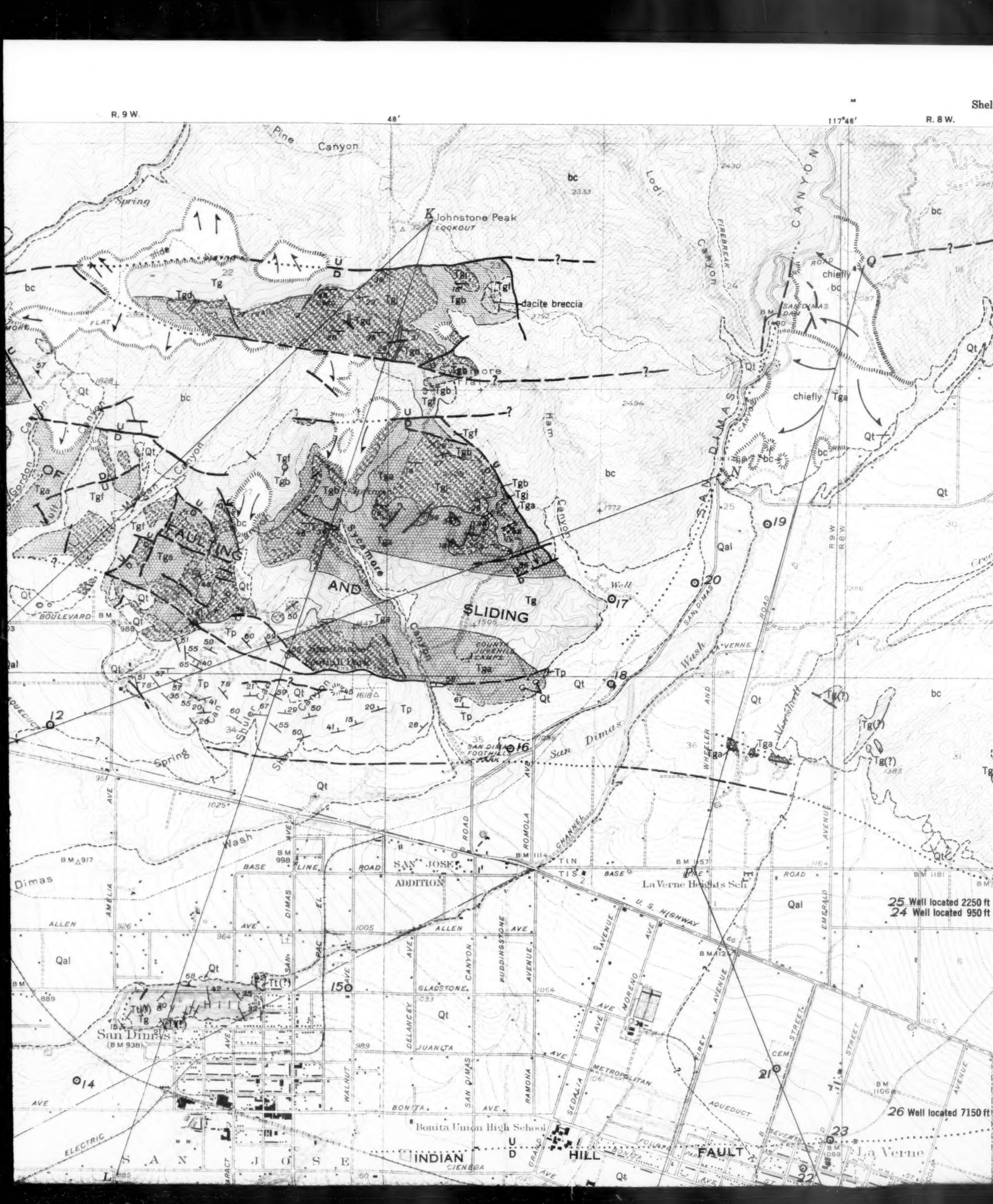
Stanford University, Stanford, California

Manuscript Received by the Secretary of the Society, July 21, 1953



GEOLOGIC MAP OF FOOTHILLS AREA NORTH OF DUARTE
(3.5 miles west of N. W. corner of larger map)

50'
R. 9 W.
Spring
Englewild
Canyon
Spring
Spring
Spring
Spring
Slide
FLAT
ZONE
Gordon
Canyon
Dalton
Wash
Little
Wash
Wash
Spring
Spring
Glendora
(B M 776)
FOOTHILL
BOULEVARD
EAST
PALM
DRIVE
AVENUE
METROPOLITAN
LIVEOAK
LORAINE
CULLEN
ATCHISON
TOPEKA
AND
AVE
AQUEDUCT
SANTA FE
HILLS
SOUTH
Dimas
San
ALLEN
BASE
LINE
DIMAS
AMELIA
HILL
CENTER
BONITA
CIENEGA
SUNFLOWER
VALLEY
LONE
ELECTRIC
AVENUE
San Dimas
(B M 938)
JOSE ADDITION
Oak School
San Dimas
Qal
Qt
Tg
Tgf
Tgd
Tp
Tt
bc
B M 776
B M 761
B M 889
B M 998
B M △917


R. 9 W.
48'
117°46'
R. 8 W.
Pine Canyon
Spring
Johnstone Peak
LOOKOUT
Lodi Canyon
FIREBREAK ROAD
SAN DIMAS CANYON
dacite breccia
chiefly
Sycamore Canyon
Ham Canyon
Gordon Canyon
Shuler Canyon
FAULTING AND SLIDING
SAN DIMAS CANYON ROAD
SAN DIMAS WASH
R 9 W R 8 W
Well
Qal
Qt
Tg
Tp
Tga
Tgb
Tgd
Tgf
Tgi
bc
Wash
San Dimas
SAN JOSE ADDITION
La Verne Heights Sch
U. S. HIGHWAY
San Dimas (BM 938)
Bonita Union High School
INDIAN HILL FAULT
SAN JOSE HILL
La Verne
25 Well located 2250 ft
24 Well located 950 ft
26 Well located 7150 ft

EXPLANATION

CENOZOIC

QUATERNARY

Recent

Qal

Alluvium

Grayish, with rounded pebbles and boulders

Pleistocene

Qt

Old alluvium

Terraces at various levels; valley-filling San Dimas formation (reddish with more angular pebbles)

TERTIARY

Pliocene (?)

Tdc

Duarte conglomerate

Well-rounded pebbles to boulders in sandy matrix

Miocene

Upper

Tp

Puente formation

Marine siltstone, sandstone and siliceous shale with tongues of conglomerate

Middle

Tt

Topanga formation

Marine sandstone, boulder-sandstone and conglomerate, locally interbedded with upper members of Glendora volcanics

Middle Miocene and possibly older

Tg

Glendora volcanics
Undifferentiated

Tgc

Volcanic conglomerates of pre-Puente age

Tgd

Dikes

Chiefly doleritic

Tga

Andesites, chiefly flows

Including hypersthene andesite breccia west of Mountain Meadows, fine-grained biotite andesite at Puddingstone Dam, hornblende andesite in Harwood Hills, and calcic andesite flows of foothills; stippled areas indicate associated bedded tuffs

Tgt

Tuff-breccia chiefly andesitic

Exposed in the Harwood Hills and south of Johnstone Peak

Tgb

Basalts

Including fine-grained basalt porphyry west of Ganesha Park, pillow lavas west of Puddingstone reservoir, and basalt flows in the foothills; stippled areas indicate palagonitic tuff west of Puddingstone dam

Tgr

Dacites and rhyolites

Including spadra felsophyre, and felsitic rhyolite of Harwood Hills; stippled area indicates Elephant Hill breccia

Tgf

Fine-grained andesite of foothills area

Pre-middle Miocene

md

Mountain Meadows biotite dacite porphyry

MESOZOIC

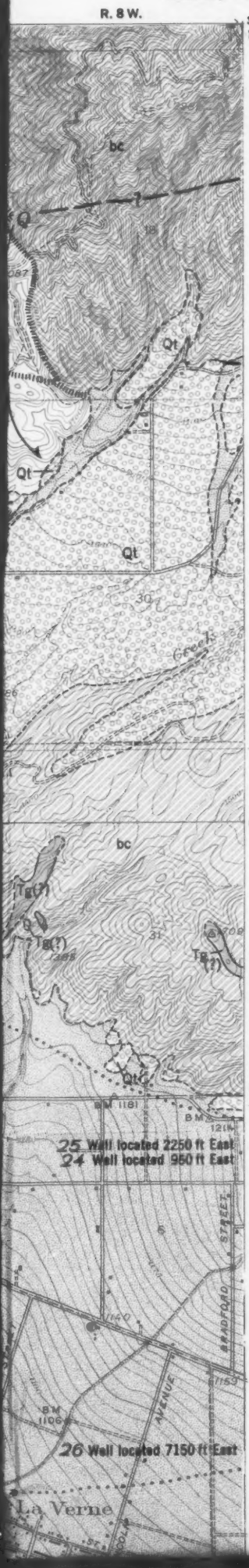
CRETACEOUS

Pre-upper Cretaceous

bc

Basement complex

Shelton, Pl. 1



EXPLANATION

CENOZOIC

QUATERNARY

Recent

Qal

Alluvium

Grayish, with rounded pebbles and boul

Pleistocene

Qt

Old alluvium

Terraces at various levels; valley-filling San
formation (reddish with more angular pe

TERTIARY

Pliocene (?)

Tdc

Duarte conglomerate

Well-rounded pebbles to boulders in sandy m

Miocene

Upper

Tp

Puente formation

Marine siltstone, sandstone and siliceous
with tongues of conglomerate

Middle

Tt

Topanga formation

Marine sandstone, boulder-sandstone a
conglomerate, locally interbedded wit
upper members of Glendora volcanic

Middle Miocene and possibly older

Tg

Glendora volcanics
Undifferentiated

Inclu
bre
fin
Pu
an
cal
sti
bed

MESOZOIC

CRETACEOUS

Pre-middle Miocene

mdp

Mountain Meadows biotite dacite porph

Pre-upper Cretaceous

bc

Basement complex

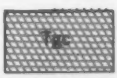
boulders

San Dimas
ar pebbles)

ndy matrix

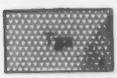
ceous shale

one and
d with
canics

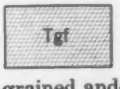


Volcanic conglomerates
of pre-Puente age

Dikes
Chiefly doleritic



Andesites, chiefly flows
Including hypersthene andesite breccia west of Mountain Meadows, fine-grained biotite andesite at Puddingstone Dam, hornblende andesite in Harwood Hills, and calcic andesite flows of foothills; stippled areas indicate associated bedded tuffs

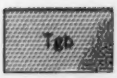


Fine-grained andesite
of foothills
area

porphyry

Tuff-breccia
chiefly andesitic
Exposed in the Harwood Hills and south of Johnstone Peak

Dacites and rhyolites
Including spadra felsophyre, and felsitic rhyolite of Harwood Hills; stippled area indicates Elephant Hill breccia



Basalts
Including fine-grained basalt porphyry west of Ganesha Park, pillow lavas west of Puddingstone reservoir, and basalt flows in the foothills; stippled areas indicate palagonitic tuff west of Puddingstone dam

GEOLOGIC MAP OF FOOTHILLS AREA NORTH OF DUARTE
(3.5 miles west of N. W. corner of larger map)

Tongues of sandstone and conglomerate

GEOLOGIC MAP OF G

Complexly mixed volcanics and sediments

PUDDINGSTONE DAM

Puddingstone Reservoir

Approximate shore line

Tgb pillow lavas

Tp (?)

Harwood Hills

Indian Hill

Bonita Union High School

San Dimas Junction

Mountain Meadows Country Club

Elephant Hill

Union Hill

Diamond Bar Ranch

SAN JOSE HILLS

SAN JOSE FAULT (Groundwater evidence)

SAN JOSE WASH

SAN JOSE CREEK

SOUTHERN PACIFIC

SOUTHERN PACIFIC

ATLANTIC AND PACIFIC

PACIFIC ELECTRIC

Scale 1/24000

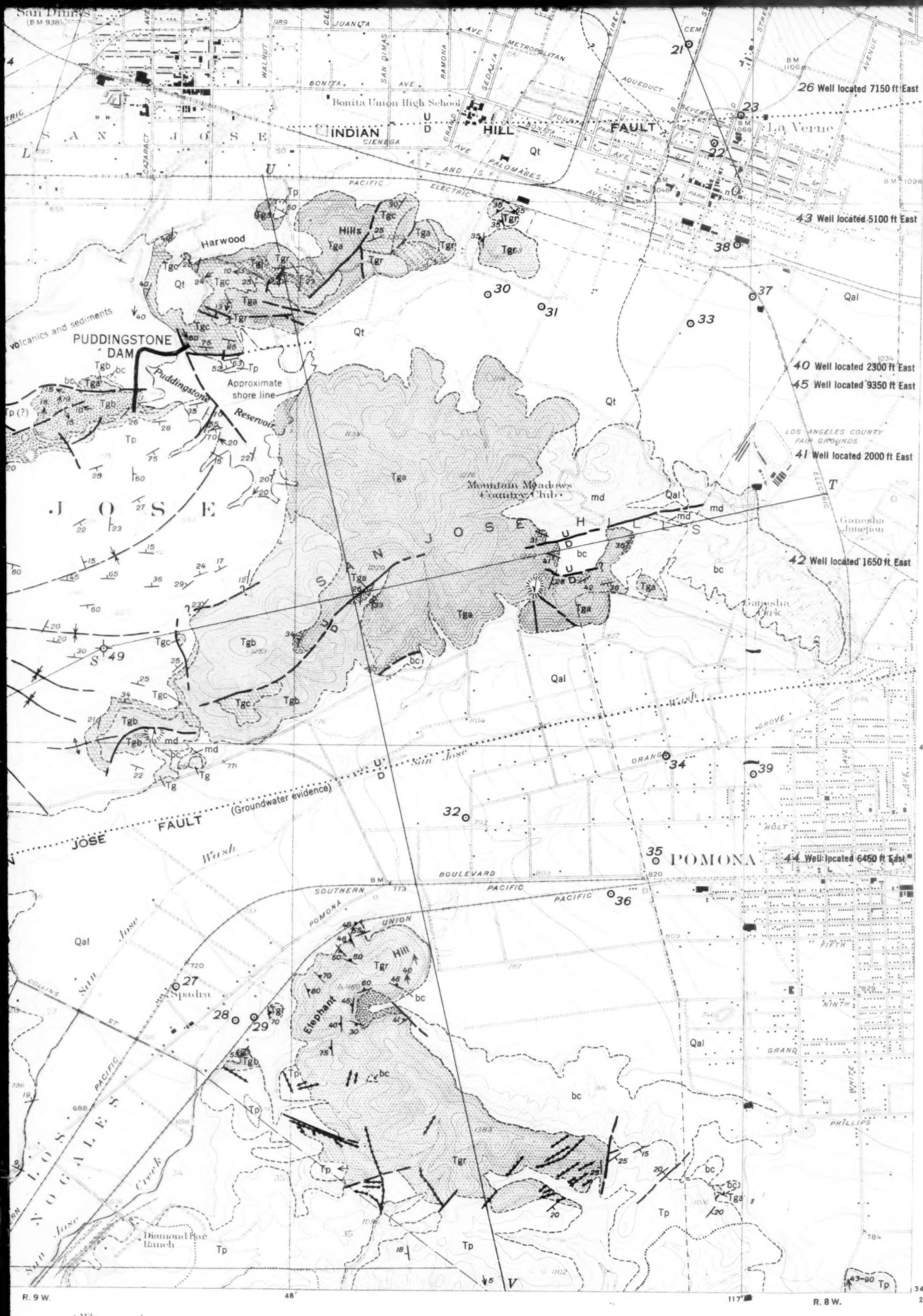
1/2 0 1 Mile

3000 2000 1000 0 5000 Feet

Contour interval 5 feet and 25 feet

Datum is mean sea level

GLENDORA VOLCANIC ROCKS, CALIFORNIA



ROCKS, CALIFORNIA

MESOZOIC

CRETACEOUS

Pre-upper Cretaceous

Pre-mid Miocene

mtd

Mountain Meadows biotite dacite porphyry

bc

Basement complex

Exposed contact

Inferred contact

Intraformational contact or bed

U
D

Exposed and inferred fault
U, *upthrown side;* D, *downthrown side*

Concealed fault

Approximate crest of anticline showing plunge

Approximate trough of syncline

25

Strike and dip of beds

84

Strike and dip of overturned beds

Strike of vertical beds

20

Generalized strike and dip of crumpled, or undulating beds

72

Apparent dip of beds or volcanics

30

Strike and dip of foliation

30

Strike and dip of foliation or flow structure in volcanic rocks

Carbonate vein

Recent landslide; boundary approximate

Water well

Corehole or well drilled for oil

Fine-grained andesite of foothills area

Including spadra felsophyre, and felsitic rhyolite of Harwood Hills; stippled area indicates Elephant Hill breccia

6′

4′

34°
2′

ROCKS, CALIFORNIA

MESOZOIC

CRETACEOUS

Pre-upper Cretaceous

Pre-mi Mioce

Mountain Meadows biotite dacite porphyry

Fine-grained andesite of foothills area

Including spadra felsophyre, and felsitic rhyolite of Harwood Hills; stippled area indicates Elephant Hill breccia

bc

Basement complex

Exposed contact

Inferred contact

Intraformational contact or bed

U / D

Exposed and inferred fault

U, *upthrown side;* D, *downthrown side*

Concealed fault

Approximate crest of anticline showing plunge

Approximate trough of syncline

25

Strike and dip of beds

84

Strike and dip of overturned beds

Strike of vertical beds

20

Generalized strike and dip of crumpled, or undulating beds

72

Apparent dip of beds or volcanics

30

Strike and dip of foliation or flow structure in volcanic rocks

Carbonate vein

Recent landslide; boundary approximate

Water well

Corehole or well drilled for oil

In
D

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 66, PP. 45-90, 3 FIGS., 6 PLS. JANUARY 1955

GLENDORA VOLCANIC ROCKS, LOS ANGELES BASIN, CALIFORNIA

BY JOHN S. SHELTON

ABSTRACT

The Los Angeles Basin in southern California is filled with locally derived Cenozoic rocks probably exceeding 25,000 feet in maximum thickness. Marginal facies of some of these sediments are exposed in and near the foothills of the San Gabriel Mountains from west of Glendora to the vicinity of Pomona, Los Angeles County, where they either rest on or are interbedded with volcanic rocks. These last, the Glendora volcanics, consist of massive and autobrecciated basalt, calcic andesite, andesite, dacite(?), and rhyolitic lavas, and an approximately equal volume of interbedded tuffs and tuff breccias. Among both flows and pyroclastics andesitic compositions predominate. Similar rocks encountered in wells in the eastern half of the Basin are considered part of the same series.

The later members of the Glendora volcanics are interbedded with middle Miocene marine sediments; the earlier representatives are probably not older than Miocene and may increase in age toward the center of the Basin. Maximum known thickness occurs in wells and exceeds 3500 feet.

The exposed Glendora volcanics here studied constitute but a small, much-faulted and eroded remnant of a volcanic field whose original extent was probably at least 35 miles from north to south and almost as much from east to west. Local evidence shows that part of the northeastern shore of the middle Miocene sea passed through the mapped area. Considerable volumes of andesitic tuff breccia containing, on the landward side, numerous large blocks with what are interpreted as radial cooling cracks are believed to be the deposits of several eruptions of *nuée ardente* type. The fact that within an 8- by 9-mile rectangle more than 20 varieties of volcanic rock are exposed in less than 8 square miles of outcrops seems to indicate a considerable number of small vents or a few remarkably versatile sources. The scene may have resembled Vesuvius and the Phlegraean Fields bordering the Bay of Naples.

Fourteen new chemical analyses of members of the volcanic series emphasize their general andesitic character and notably low K_2O and ferromagnesian constituents.

CONTENTS

TEXT

	Page
Introduction	46
Descriptive geology	47
General statement	47
Basement complex	53
General statement	53
Mountain Meadows dacite	54
Age of basement complex	54
Glendora volcanics in general	54
Glendora volcanics east of Spadra	55
Location	55
Spadra felsophyre	55
Elephant Hill breccia	56
Other volcanics	56
Glendora volcanics of the Ganesha area	56
Location	56
Hypersthene andesite breccia	57
Fine-grained basalt porphyry	57
Glendora volcanics of the Puddingstone area	58
Location	58
Early tuffs and basalts	58
Fine-grained biotite andesite	59
Felsitic rhyolite	59
Perlite	59
Hornblende andesite	60
Tuff breccia	61
Palagonitic tuff and pillow lava	61
Volcanic conglomerates	61
Area of complexly mixed volcanics and sediments	62
Glendora volcanics in Way Hill and Lone Hill	62
Glendora volcanics in the Glendora South Hills	63
Glendora volcanics in the Foothills area	63
General statement	63
Fine-grained andesite	64
Calcic andesite flows	65
Johnstone Peak tuff breccia	65
Basaltic lava	67
Hornblende andesite (tuff) breccia	68
Dikes and other possible intrusive masses	68

	Page
Inter-area correlation of the Glendora volcanics	69
Petrology and chemical composition of the Glendora volcanics	70
Topanga formation	71
Puente formation	76
Duarte conglomerate	76
San Dimas formation and terrace deposits	77
Structure	77
Geomorphology	79
Age of the Glendora volcanics	79
Interpretative geology	80
Extent of probable correlatives of the Glendora volcanics	80
Subsurface	80
Surface	80
Probable volume	81
The source problem	81
Summary: The volcanic episode reconstructed	88
References cited	88

ILLUSTRATIONS

Figure	Page
1. Map of a portion of coastal southern California	48
2. Diagrammatic columnar section for area covered by Plate 1; dikes not shown	53
3. Niggli-Becke-Bacon diagram comparing chemical compositions of Glendora volcanics with Tröger's types and Daly's averages	75
Plate	**Facing page**
1. Geologic map of Glendora volcanics, California	45
	Following page
2. Photomicrographs of Glendora volcanics south of the foothills area	66
3. Photomicrographs of Glendora volcanics from the foothills area	66
4. Johnstone Peak tuff breccia	66
5. Aerial view of Glendora South Hills and foothills of San Gabriel Mountains	66
	Facing page
6. Structure sections	90

TABLES

Table	Page
1. Summary of information concerning volcanics as reported from wells shown in Figure 1	49
2. Chemical analyses, C.I.P.W. norms and Niggli values	72
3. Wells and coreholes affording reliable information on suballuvial geology within area covered by Plate 1	82

Introduction

In Southern California the term Los Angeles Basin is used in a geologic sense for a restricted area of Cenozoic subsidence and the thick sequence of locally derived sediments with which it is filled. As now exposed this structural depression is roughly 40 by 60 miles, elongate northwestward, and centered about 15 miles southeast of the city of Los Angeles. In the central portion there are no outcrops of the rocks beneath the alluvial veneer; the deepest wells at present do not penetrate below the upper Miocene, and estimates of total thickness range upward from 25,000 feet. However, along some of the margins, especially to the north and east, there are scattered areas, which, taken together, afford the most nearly complete known sample of the basin deposits. Thus the best outcrop section of Paleocene, Eocene, Oligocene(?), and lower and perhaps middle Miocene rocks is in the northern Santa Ana Mountains (Fig. 1). Middle and upper Miocene are well exposed in the Santa Monica Mountains, upper Miocene in the Puente and San Jose Hills, and Pliocene in the Repetto and adjacent western Puente Hills. The remarkable marine Pleistocene of the Palos Verdes Hills is probably familiar to most readers (*see*, for example, Woodring *et al.*, 1946). Compared to subsurface data from out in the Basin, these somewhat deformed fringes commonly exhibit different thicknesses, coarser facies, and local unconformities suggesting fluctuating shore lines and sources of supply.

At many places around the periphery of the Los Angeles Basin (Fig. 1) exposed extrusive rocks are intimately associated with Miocene marine sediments. One of the most extensive of these, and possibly the most complex petrographically, is the west-to-east arc from west of Glendora to south of Pomona throughout which marginal facies of the basin sediments either rest on or are interbedded with volcanic rocks. These last, the Glendora volcanics (Shelton, 1946), consist of massive and autobrecciated lavas, tuffs, and tuff breccias, largely andesitic but ranging from olivine basalt to rhyolite. Similar rocks encountered in wells in the eastern half of the Basin are considered part of the same series. Although the name, as here used, is intended only for the volcanic rocks in the northeastern part of the Basin, the strong probability that the volcanics of other parts are essentially contemporaneous suggests that future work may justify extension of this or a more appropriate term.

This paper presents the results of detailed field mapping of the Glendora volcanics in their type area combined with petrographic study and chemical analyses of both outcropping material and well cores. For practical reasons, Plate 1 carries less than half the number of units recognized in the field, and these at less than one third the field map scale. This consolidation has obscured some of the small-scale structure, and the result gives an unwarranted impression of similarity of units and of correlation from one outcrop area to another. The variety thus lost is indicated in the text and retained in the lithology and contacts shown on the structure sections (Pl. 6).

The work has been carried on intermittently since 1944. Grateful appreciation is expressed to A. O. Woodford, C. R. Longwell, Adolph Knopf, and C. A. Anderson for general counsel, to numerous representatives of locally operating oil companies for helpful discussions and data on the subsurface geology, to the Director of the U. S. Geological Survey for permission to use data obtained while in its employ, to M. N. Bramlette, M. L. Natland, and W. T. Rothwell, Jr., for study of Foraminifera, and to Dr. L. R. David for her work on many fish-scale collections. Some of the subdivision of the Quarternary is taken from unpublished maps by Rollin Eckis. The Duarte fault west of Azusa is based partly on an unpublished report submitted to the Los Angeles County Flood Control District in 1931 by the late Willis S. Jones, an engineer.

The chemical analyses were made possible by a research grant from the Penrose Bequest of The Geological Society of America whose cooperation in this and other matters is gratefully acknowledged.

Other than that already referred to, there is no published work on the volcanic rocks as such. Five reports, with maps, include all or part of the area here considered: Eldridge (1907) named the Puente formation; Mendenhall (1908) worked primarily with the Quaternary deposits; English (1926) made the first general geologic study of The Puente and San Jose Hills, but his work was done just too early to benefit from the use of microfossils; Eckis (1928; 1934) in his report on the Quaternary deposits and ground-water conditions set a new standard for such studies in this region; Woodford *et al.* (1944) presented a map which is a compilation of published and unpublished data, utilizing Foraminifera. Part of the area is included in a report by the same authors (Woodford *et al.*, 1946) on the conglomerates of the region.

The first report on the volcanic rocks as such is an unpublished thesis by Bert H. Mull (1934); for the sake of uniformity of treatment his area was remapped by the present author. The only other work in which the volcanic rocks are systematically described is the author's preliminary map of the northeast margin of the San Gabriel basin (Shelton, 1946).

Descriptive Geology

General Statement

The portion of the Los Angeles Basin here considered (Pl. 1) is characterized by marked basinward thickening and changes in facies. The pre-Upper Cretaceous crystalline rocks of the basement are extensively exposed and have been reached in some wells. They are overlain by from a few hundred to over 3000 feet of flows, breccias, and tuffs belonging to the Glendora volcanic series (Fig. 2). The upper part of these is in some places interbedded with marine sandstones, boulder sandstones, and conglomerates of the middle Miocene Topanga formation which is about 2000 feet thick in this area. More commonly the volcanics are unconformably overlain by some part of the extensive upper Miocene Puente formation which is composed chiefly of siltstones and shales with minor sandstones. In the foothills these are approximately 2500 feet thick; south and southwest of the mapped area they reach 9000 feet. Although over most of the area these constitute the youngest exposed Tertiary rocks, in the western part of the foothills belt, north of Duarte, a coarse boulder conglomerate probably at least 1500 feet thick is believed, on indirect evidence, to be Pliocene. This Duarte conglomerate (Shelton, 1946) may be the terrestrial equivalent of part of the thick marine Pliocene section exposed 11 miles to the southwest where lower Pliocene sediments alone reach 5000 feet. The Pleistocene San Dimas forma-

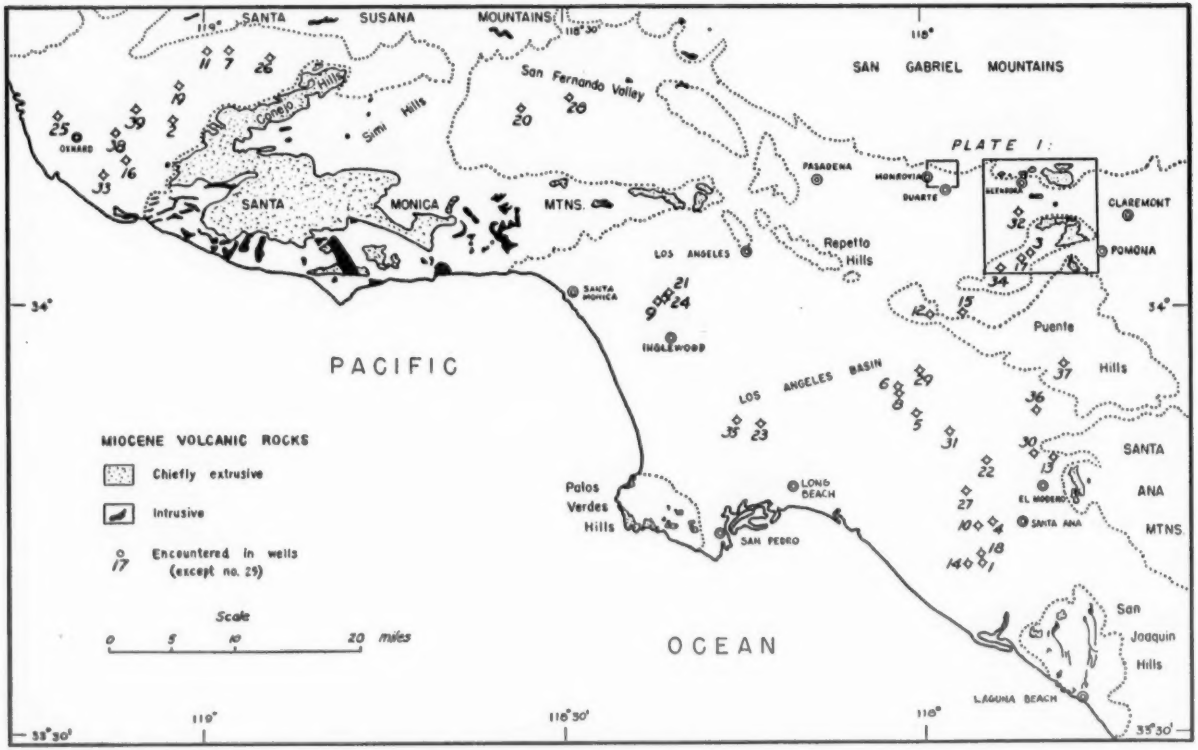


FIGURE 1.—MAP OF A PORTION OF COASTAL SOUTHERN CALIFORNIA

Showing location of area studied (Plate 1), outcropping Miocene volcanics, and wells in which (except No. 25) volcanic rocks have been encountered; *see also* Table 1.

TABLE 1.—SUMMARY OF INFORMATION CONCERNING VOLCANICS AS REPORTED FROM WELLS SHOWN IN FIGURE 1

TABLE 1.—SUMMARY OF INFORMATION CONCERNING VOLCANICS AS REPORTED FROM WELLS SHOWN IN FIGURE 1

Map No.	Operator	Well	Elevation of top of volcanics	Known penetration (in feet) and character of volcanics	Volcanics reported overlain by	Volcanics reported underlain by
1	Bandini Petrol. Co.	Segerstrom 1	−5,539±	1,300+ (?) Hypersthene augite basalt at least in part; much tuff breccia, prob. andesitic	?	Bot. in vol. (?)
2	Calif. Petrol. Corp.	Arneil 1	−1,000	1,273+ Basalt and basalt breccia	Low U. Mioc.	Bot. in vol.
3	Continental Oil Co.	Bayly 1	−1,023±	1,765± Probably chiefly andesite, largely fragmental. (*See also* Table 3)	M. Mioc. Topanga fm.	Basement
4	Continental Oil Co.	Santa Ana Comm. 1	−4,130±	200+ Altered basalts interbedded with M. Mioc.	"Pico"	Bot. in vol.
5	Gen. Petrol. Corp.	Heath 1	−11,062	288+ Basalt interbedded with M. Mioc. foraminifera	L. Pliocene Repetto fm.	Bot. in vol.
6	Gen. Petrol. Corp.	La Mirada 46-1	−12,480±	47+	High M. Mioc.	Bot. in vol.
7	Gen. Petrol. Corp.	Las Posas-McBean 1	−3,052	770 Chiefly basalt breccias. Some andesite, vesicular flows and bentonite	Low U. Mioc. Monterey fm.	Vaqueros sandstone
8	Gen. Petrol. Corp.	Librown 1	−12,000(?)	600± Interbedded sediments and volcanics to bottom	U. Mioc. Puente fm.	M. or U. Mioc.
9	Gen. Petrol. Corp.	Sentous 1	−8,049	62 Probably basalt; altered; fine grained. Some interbedded sediments reported M. Mioc.	U. Mioc.	M. Mioc.
10	Gen. Petrol. Corp.	Was 1	−5,598±	160+ Basalt (?) breccia	M. Mioc. (?)	Bot. in vol.
11	Humble Oil and Ref. Co.	Berylwood 1	−2,294	806 Chiefly basalt breccias. Some andesite, vesicular flows and bentonite	Low U. Mioc.	Sespe (unconformably)

TABLE 1.—*Continued*

Map No.	Operator	Well	Elevation of top of volcanics	Known penetration (in feet) and character of volcanics	Volcanics reported overlain by	Volcanics reported underlain by
12	Los Angeles Brewing Co.	Jones 1	+305±	1000+ Altered tuffs and tuff breccia, largely palagonitic	Low U. Mioc.	Faulted against lower Puente or older siltstones
13	McKee Oil Co.	Kokx Comm. 8-1	−1,871	Unknown thickness; altered volcanic rock	?	*Turritella ocoyana*
14	Morton and Sons	Crew Comm. 1	−8,850±	850±	U. Mioc.	Bot. in vol.
15	Morton and Sons	Rowland Ranch Estate 3-1	−7,140±	980± Diabasic basalt. (209± of diabase higher in the hole)	M. Mioc. (?)	Granite (?) etc.
16	Mugu Syndicate	Utt 1	−3,977	298+ Basalt 319+ Fragmental basalt with some andesite	High Plioc. (unconf.)	Topanga sh and ss (probably interb'd)
17	The Ohio Oil Co.	Legrand 1	−1,742	2,990± Chiefly andesite; some basalt. (*See also* Table 3)	M. Mioc. Topanga fm.	Basement
18	P. M. Girard	Mark Fisher 1	−5,165	450+ Fragmental andesites (?)	Repetto	Bot. in vol.
19	Richfield Oil Co.	McFarland 1	−553	2,335 Altered, vesicular basalt and andesite(?) breccias and tuffs; possibly some interbedded sediments	Monterey fm.	Vaqueros fm.
20	Richfield Oil Co.	Mulholland 1	−3,560	400 Basalt, massive	M. Mioc. Topanga fm.	M. Mioc. Topanga fm.
21	Shell Oil Co., Inc.	Baldwin Hills Comm. 1	−13,022	158+ Andesite, near basalt	M. Mioc.	Bot. in vol.

22	Shell Oil Co., Inc.	Mathis 1	−4,714	1,074+ Basalt and tuff, with some admixed sand	U. or M. Mioc.	Bot. in vol.
23	Shell Oil Co., Inc.	Reyes 135	−10,852	1,106 Fine grained andesite and/or basalt; largely fragmental	U. Mioc.	Basement
24	Standard Oil Co., Calif.	Baldwin Cienega 105	−7,733	About 38% of lowest 4180 feet is volcanics	M. Mioc.	Bot. in vol. or basement
25	Standard Oil Co., Calif.	Eastwood 1	No volcanics; entered Sespe at −9,345±; T.D. 11,027)			
26	Standard Oil Co., Calif.	Gabbert 1	−1,850??	Volcanics on ditch, mostly basalt	?	Bot. in Sespe
27	Standard Oil Co., Calif.	Stanley Comm. 1	−8,000±	424+	Repetto or older (?)	
28	Standard Oil Co., Calif.	Woo 1	−7,799	125 Fragmental altered basalt	Mohnian(?)	M. Mioc. Topanga fm.
29	Standard Oil Co., Calif.	Woodward Comm. 1	−10,979±	196±	U. Mioc.	M. Mioc. or Sespe
30	Standard Oil Co., Calif.	Zaiser-Brelje	−4,134	100+ Probably interbedded with sediments	U. or M. Mioc.	Diabase, or ?
31	The Texas Co.	Buena Park U-1-1	−8,763	182+ Fragmental, at least in part	Mioc.(?)	Bot. in vol.
32	The Texas Co.	Covina 27-1	−365±	3,660± Andesite and basalt, fragmental and massive. (*See also* Table 3)	M. Mioc. Topanga fm.	Bot. in vol. or basement
33	The Texas Co.	Eastwood 1	−1,897	750+ Basalt and basalt breccia	Pliocene	Bot. in vol.
34	The Texas Co.	Garnier 1	−1,880	3,720± Basalt and andesite; spilite(?) (*See also* Table 3)	M. Mioc.	Mylonitic basement
35	Union Oil Co. of Calif.	Callender 79	−11,257	455± Some interbedded platy shale	U. Mioc.	U. Mioc. or M. Mioc.

TABLE 1.—*Concluded*

Map No.	Operation	Well	Elevation of top of volcnics	Known penetration (in feet) and character of volcanics	Volcanics reported overlain by	Volcanics reported underlain by
36	Union Oil Co. of Calif.	Chapman 29	−7,617±	195±	U. Mioc. Puente fm.	M. Mioc.(?)
37	Union Oil Co. of Calif.	Gaines 1	−5,169±	20+ Andesite(?). 552± of diabase higher in the hole	M. Mioc.	M. Mioc.(?) cg-ss.
38	Vaca Oil Co.	Livingston 1	−3,855	370+ Basalt and basalt breccia with streaks of ss near the base	U. Mioc.	Bot. in vol.
39	Western Gulf Oil Co.	Beardsley 1	−2,585	2,264± Basalt and andesite flows, breccia, and tuffs	Basal Monterey fm.	Temblor M. Mioc. with diabase intrusions

tion, consisting of up to several hundred feet of weathered alluvial gravels and their decomposition products, occurs as remnants of relatively flat-lying terrace deposits in many areas, although in some places close to the frontal faults of the San Gabriel Mountains it has southerly dips as high as 83°.

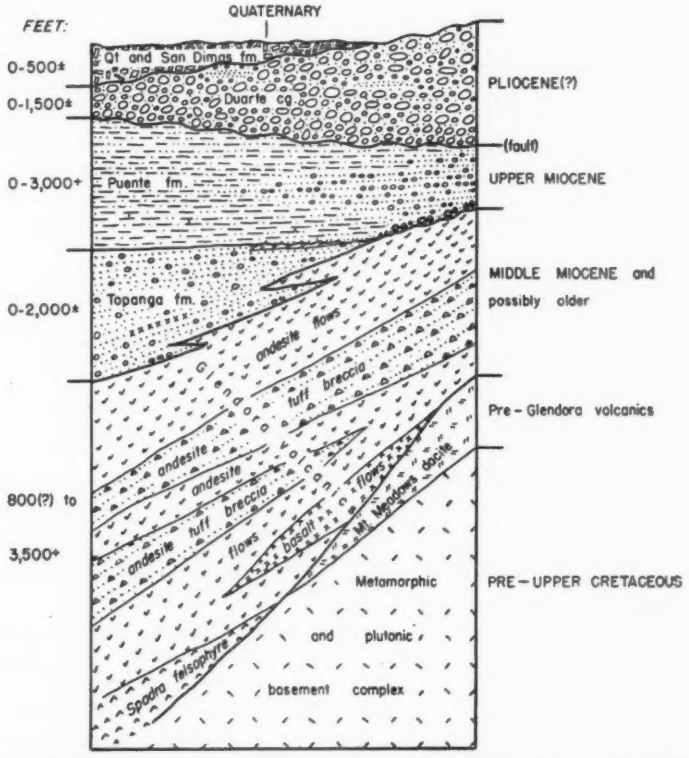


FIGURE 2.—DIAGRAMMATIC COLUMNAR SECTION FOR AREA COVERED BY PLATE 1; DIKES NOT SHOWN

Basement Complex

General statement.—Under this heading are grouped the crystalline rocks, chiefly plutonic and metamorphic, so extensively exposed in the San Gabriel Mountains and found at numerous points beneath the Tertiary deposits of the margins of the Los Angeles Basin. With one exception (the Mountain Meadows dacite) no attempt is made here to subdivide the complex, but its general character can be briefly stated. (*See* Miller, 1934; Oakeshott, 1938; Alf, 1948, for further details.)

In the south face of the San Gabriel Mountains the predominating rocks are granitic to quartz dioritic gneisses whose planar structures, in this area, generally dip steeply northward. These are cut by locally abundant pegmatite and aplite dikes and rarer dacite and diabase dikes. East of San Dimas Canyon there is a large body of quartz diorite. The gneisses south and southeast of this contain zones of mylonite which develop into bands totalling several thousand feet thick as they are traced across the face of the mountains for 18 miles east of the area of Plate 1 (Alf, 1943; 1948). Less abundant are mica schists and remnants (chiefly east of the mapped area) of quartzite, marble, and associated metasediments.

The only other extensive exposures of the

basement complex are southwest and northwest of Pomona. The former is composed of a coarse-grained biotite granodiorite and biotite-quartz diorite with associated pegmatites and aplite dikes. The rock disintegrates easily and is weaker topographically than the volcanics and sediments. Northwest of Pomona the prevailing rocks are pinkish gneissoid granite and granodiorite with associated dikes and basic inclusions and the largest exposure of a distinctive rock—the Mountain Meadows dacite.

Mountain Meadows dacite.—This biotite dacite porphyry, which well data suggest may be considerably more extensive underground than at the surface, takes its name from the outcrop area around the Mountain Meadows County Club (Woodford *et al.*, 1946, p. 527) where it is usually deeply weathered and topographically weak. Hand specimens are characterized by euhedral phenocrysts of biotite and less easily recognized oligoclase and rounded quartz grains in a pale fine-grained groundmass. One of the freshest samples so far obtained came from about 4000 feet in the Continental Oil Company's Bayly No. 1 well (No. 48 on Pl. 1; *see also* Pl. 6, section RST and Tables 2 and 3). Dikes of similar but finer-grained biotite dacite cut both the main mass in the type area and the adjacent granite, and somewhat similar rock occurs in dikes in the San Gabriel Mountains.

The texture suggests either a shallow intrusive or an extrusive origin. No vesicles, brecciation, or flow structure are known, and the highly uniform character of the rock in all outcrops and in wells extending 5 miles southwest of the Country Club further favors a shallow sill-like origin. In the Ohio Oil Company Legrand No. 1 well 0.8 mile southwest of Buzzard Peak (No. 47, Pl. 1) the Mountain Meadows dacite was penetrated (probably with low dips) for approximately 282 feet between Glendora volcanics above and granite and granite gneiss below.

Age of basement complex.—The oldest sedimentary rocks resting on the basement complex in this area are the Upper Cretaceous sandstones and shales of the northern Santa Ana Mountains (Popenoe, 1942). Here Larsen (1948, p. 136) and others have shown that the plutonic rocks are probably at least post-Triassic; if they may be correlated with similar plutons in the Klamath Mountains of northwestern California (Hinds, 1934) or in northern Lower California (Böse and Wittich, 1913), their age is probably close to early Cretaceous (post-Portlandian and pre-Turonian). (*See also* Woodford, 1939).

If the Mountain Meadows dacite is intrusive, it must be older than the Glendora volcanics, for nowhere is it known to penetrate them, whereas numerous pebbles of the dacite have been found at two places in the basal members of the volcanics as well as less abundantly in conglomerates of the Topanga and Puente formations (Woodford *et al.*, 1946). Present evidence limits its age to the range from possibly slightly ealier than the associated early Creataceous(?) plutons to pre-middle Miocene.

Glendora Volcanics in General

The term Glendora volcanics has been proposed for

> "the series of flows, breccias, tuffs, and related intrusives exposed in the foothills of the San Gabriel Mountains in the vicinity of Glendora, in the South Hills, and at the northeast end of the San Jose Hills" (Shelton, 1946).

These exposures in the type area aggregate about 7.3 square miles and indicate an estimated volume of at least 1 cubic mile. Well data indicate a subsurface extent of probably at least 300 square miles on the east side of the Los Angeles Basin; even if these were only a few hundred feet thick there would be tens of cubic miles. The volume may be much more, for in three wells on Plate 1 the volcanics were 3000 feet or more in apparent thickness.

The Glendora volcanics include undisturbed tuffs, reworked tuffs, tuff breccias, massive flows, autobrecciated flows, dikes, probable vent fillings of various kinds, volcanic conglomerates, and possibly local fumarolic alteration products. As might be inferred from Table 2, where the silica percentages range from 47.23 to 75.50, the volcanics include olivine basalt, a considerable variety of andesites, and several rocks in the dacite-rhyolite group. In the field 22 mappable varieties or members of the Glendora volcanics were recognized.

Two modifications of treatment stem from this variety: (1) As already pointed out, the 22 map units have here been reduced to 8. (2) Because of the correlation problem, it seems wiser to describe the rocks by geographic areas of outcrop rather than in an attempted overall chronologic sequence; after this, the evidence for their age and possible inter-area correlations will be discussed for the region as a whole. The occurrence and petrography of the volcanics in the six outcrop areas will be considered in order from south to north (Pl. 1).

Glendora Volcanics East of Spadra

Location.—The isolated area of volcanics exposed east of Spadra and 2 miles southwest of Pomona is referred to as the Spadra area. Here, along the ridge that forms the northern tip of the Puente Hills, the volcanics are exposed over an area of three-fourths of a square mile. They rest on plutonic basement rocks wherever their base can be seen and are overlain by conglomerates of the middle part of the Puente formation or by alluvium.

Spadra felsophyre (included in Tgr on map).—Megascopically the lavas of Elephant Hill and the ridge east of it are very fine-grained laminated flows and flow breccias. Silicification and pyritization are so extensive that no sample completely free of secondary minerals was found. Much of the rock, especially at the southeast end of the ridge, is so thoroughly altered and variegated in color that hand-lens identification is very difficult. On fresh fracture the least-altered material is usually dark gray, against which background can be recognized occasional 1- to 2-mm phenocrysts of plagioclase and scattered minute grains of pyrite. Some of the blocks in the breccias are traversed by narrow (less than 1-inch) dikelets of microbreccia composed of small angular fragments of the lava (whose flow texture is often oblique to that of the main block) surrounded by chilled groundmass and more or less chalcedony, thus establishing two stages of brecciation. Pyrite is present in both parts of the rock.

The breccia facies is well exposed in the old quarry at the east base of Elephant Hill where its crude bedding is approximately parallel to the flow laminae near by.

The breccias are interpreted as autobreccias. Judging from present exposures, they constitute at least 80 per cent, and the laminated flows not over 20 per cent of this member of the Glendora volcanics.

Petrographic Description. Both flow and breccia facies are a notably uniform dacitic(?) felsophyre. Percentages estimated:

2%	Plagioclase, An_{20}–An_{35}, in 0.2 to 2.0 mm laths
30–50	Same, under 0.2 mm
50+	Groundmass glass, n close to 1.52, somewhat devitrified and charged with feldspar microlites in fluidal arrangement; local patches of crystobalite(?)
5	Opaques, chiefly secondary pyrite, in places suggesting former hornblende

Scattered minute birefringent particles indicate trace of some ferromagnesian mineral.

Occasionally the flows contain euhedral red garnets less than 2 mm in diameter. The most likely source seems to be the local basement complex (pale biotite-quartz diorite) which contains some banded pegmatites with similar garnet crystals.

Along the basal contact of the felsophyre where it rests on basement complex east of Elephant Hill there are local lenses of limestone up to perhaps 40 feet thick. In places these show poorly defined concentric structures or contain small fragments of altered volcanic rock and flakes of biotite. The upper contact of the volcanics is sharp and shows almost no evidence of channeling. The overlying Puente conglomerates contain a great variety of pebbles among which volcanic rocks are conspicuously rare. The maximum thickness of the exposed Spadra felsophyre is about 350 feet (Pl. 6, section UV).

In the region of exposure of this upper contact a number of vertical carbonate veins cut the volcancis and in a few cases the overlying sediments. They range from 1–2 feet to 15–20 feet wide; the majority are 4–10 feet. Two of them can be followed for about 1700 feet horizontally. They are filled with cream-colored to dark-brown limestone and magnesian limestone with local pods of coarsely crystalline cleavable carbonate and fragments of the felsophyre. The largest, in the western part of the area, has been quarried, and the cuts reveal some concentric mammillary structures a few

inches across. Evidently in post-lower upper Miocene time local warping produced fractures under such ground-water conditions that carbonates of unknown origin migrated into them in solution. This hypothesis is supported by the fact that the veins occur predominantly in northwest- and northeast-trending sets, and a few occupy faults along which the base of the sediments has been offset.

Elephant Hill breccia (stippled Tgr on map).—This member is confined to a steep-walled semicircular valley head 500 feet in diameter that penetrates the east flank of Elephant Hill. This depression is floored with a breccia composed of angular fragments of laminated gray spherulitic glass in a tuffaceous matrix. Most of the blocks are an inch or less across, but the largest are at least 2 feet long. While the glassy blocks are the only kind seen in most outcrops, there are places, such as near the southwest boundary of the breccia, where as many as six rounded cobbles of basement rocks are present per square yard of exposure. These range up to about 6 inches across and represent a variety of coarse-grained and aplitic biotite plutonites. Much rarer are blocks of Spadra felsophyre. Locally the matrix is rich in minute biotite flakes.

Petrographic Description. Blocks from Elephant Hill breccia. Percentages estimated:

20–25% Plagioclase, euhedral to subhedral, chiefly albite less than 0.1 mm, but some 2 mm oligoclase
1–2 Quartz, subhedral
1–2 Biotite, in shreds
75± Groundmass of glass containing a variety of essentially isotropic particles in long, irregular and droplike forms.

Some specimens spherulitic. Usually finely layered, many laminae crenulated. Some feldspar laths show slight bending and optical strain. A pyroclastic rock modified in the vicinity of a vent?

Chemical analysis: Table 2 (S-673). Photomicrograph: Pl. 2, fig. 1.

The possibility that the Elephant Hill breccia marks an eroded vent in supported by details of lithology and structure as well as the fact that in the San Jose Hills the volcanics are overlain by lower Puente shales whereas at Elephant Hill they are overlain by middle Puente sandstones and conglomerates. This means that in the little over a mile separating the two areas (concealed by San Jose Wash) something of the order of 200 feet of sediments has lapped out, indicating a topographic high near Elephant Hill in the early late Miocene. A water well (Pl. 1, No. 28) at the west base of the hill encountered a glassy breccia very similar to that filling the "vent". This may be some of the extruded material.

Other volcanics (included in Tg and Tga on map).—At the east end of the Spadra area there is a small outcrop of fine-grained biotite andesite petrographically similar to the more extensive exposures north and west of Puddingstone Reservoir. Soil obscures its relations to surrounding rocks, but it probably rests on or is faulted against basement complex and overlain by Puente sediments.

At the west end of the area, at the edge of San Jose Wash, there is a small quarry in a fine-grained basalt and associated bedded biotite tuff which are overlain by Puente sediments.

Glendora Volcanics of the Ganesha Area

Location.—In the northeastern end of the San Jose Hills, south and east of Puddingstone Reservoir, the Glendora volcanics are exposed over an area of 1.7 square miles. The outcrops form an isolated patch limited to the north and south by alluvium, whereas to the east they have been eroded from the basement complex of Ganesha Park and to the west they are overlain by shales of the lower part of the Puente formation.

This westward-dipping sequence consists essentially of up to 1000 feet of hypersthene andesite breccia succeeded by almost as much fine-grained basalt. Considerable amounts of tuff are associated with both (Pl. 6, section RST).

The main body of hypersthene andesite breccias rests on basement complex or Mountain dacite along most of the exposed contact, but locally up to about 75 feet of earlier breccias, flows, and more or less reworked tuffs is intercalated at this horizon (stippled Tga on map). In places, fragments of basement complex and Mountain Meadows dacite are present in these early accumulations in which seven of the

eight units recognized are andesitic. Where discernible, the relief on the pre-volcanic surface is very low.

Hypersthene andesite breccia (included in Tga on map).—The most extensive member of the Glendora volcanics, in terms of present-day continuous outcrops, is the succeeding hypersthene andesite breccia exposed west of Mountain Meadows Country Club. It covers a little over 1.1 square miles with relatively high uniformity of lithologic character. The fresh rock from blocks in the typical breccia is a dark-gray porphyry composed of phenocrysts of glassy plagioclase and less common euhedral pyroxenes in an aphanitic groundmass. In hand specimens the number of pyroxene phenocrysts ranges from about one in 5–10 square inches to two or three per square inch of surface. Vesicles are generally absent. In some places glomeroporphyritic clots of pyroxenes and plagioclase about an inch across are numerous. Most outcrops disclose a breccia composed of angular blocks of such rock averaging less than 1 foot across but reaching maxima of 4 feet or more in a matrix of similar material whose lower resistance to weathering tends to leave the blocks standing in relief. Most weathered surfaces are reddish or purplish brown from oxidation, and further variations in color in the interiors of blocks suggest alteration during their cooling history. Massive lava with crude nearly vertical columnar jointing was seen in the canyon southeast of summit 1220, and a consolidated tuff breccia can be seen in some outcrops, but most of the unit is interpreted as flow breccia.

Petrographic Description. Micrometric analysis of two thin sections:

	S-468	*S-468-2*	*Average*
Labradorite (An_{50-60})	33.0	31.6	32.3
Hypersthene	4.9	4.5	4.7
Augite	1.8	1.6	1.7
Groundmass (incl. magnetite and minute feldspars)	60.3	62.3	61.3
	100.0	100.0	100.0

Plagioclase in 2–3 mm phenocrysts, fresh, euhedral; generally showing oscillatory zoning, more sodic toward margins.

Hypersthene in 1–2 mm phenocrysts, euhedral or subhedral, commonly altered along boundaries and cracks; pleochroism weak, X = pale pinkish yellow, Z = pale greenish blue; (—) 2V close to 80°, probably about 25 molecular percent $FeSiO_3$ (Winchell, 1951, p. 406).

Groundmass contains minute feldspars, scattered magnetite and ragged patches of micropoikilitic quartz in a glassy base with n well below 1.54. Regarding similar textures, see Iddings (1899, p. 132–133), Lacroix (1908, p. 51–58) and Geijer (1913, p. 51–80).

Chemical analysis: Table 2 (S-468). Photomicrograph: Pl. 2, fig. 2.

The thickness of the exposed hypersthene andesite is probably a maximum south of Puddingstone Reservoir, where it is inferred to be about 800 feet (Pl. 6, section RST).

Several lenses of bedded gray granular hypersthene andesite lithic tuff are associated with the flow breccias. They have moderate westerly dips and are composed of fragments of fine-grained glassy volcanic rock 60–80%, labradorite crystals (An_{50-60}) 10–15%, hypersthene crystals 3–5%, rare augite, and glass 5–15%. These are similar to some of the earlier tuffs described above and are considered reworked ash falls probably closely associated with the flow breccias.

Fine-grained basalt porphyry (included in Tgb on map).—Between the hypersthene andesite breccias and the Puente marine sediments is a westward-dipping series of darker and finer-grained lavas. The fresh rock is black and usually composed of scattered plagioclase and pyroxene phenocrysts in an aphanitic groundmass (Pl. 2, fig. 3). Color variations include reddish tones produced by oxidation, especially along fractures and boundaries between flow laminae, and bleaching to pale greenish and grayish, evidently produced by deuteric alteration. Flow structure is well developed in some places in the form of slightly irregular laminations a few millimeters thick along which the rock tends to split. Elsewhere the rock is an autobreccia of subangular to angular blocks, many somewhat platy like the above, in a more or less crystalline matrix. Except for the zone near its basal contact, the rock resists weathering about as well as the hypersthene andesites.

Petrographic Description. Percentages estimated:

5–15% Plagioclase phenocrysts, An_{70-40}, commonly in clusters; zoning less prominent than in andesites

5–10 Hypersthene and augite phenocrysts, approximately equal abundance, rounded to subhedral, under 5 mm

75+ Groundmass, pilotaxitic to hyalopilitic (glass 1–15%) with clinopyroxene, magnetite, and feldspars in striking fluidal arrangement.

Photomicrograph: Pl. 2, fig. 3.

The best exposures of this member are in steep canyons and road cuts along its south edge where a small part (probably less than 25 per cent) consists of local accumulations of tuffs, tuff breccias, and tuffaceous breccias. Maximum exposed thickness is probably at least 500 feet. The upper surface was subjected to erosion and some channelling before deposition of the overlying Puente sediments. On Plate 1 the volcanic conglomerates along this contact are grouped with the volcanic conglomerates of the Puddingstone area, under which heading they are discussed.

Glendora Volcanics of the Puddingstone Area

Location.—Under this heading are grouped the volcanic rocks northeast and southwest of Puddingstone Dam. The total area of exposed volcanics is about 1.25 square miles and is the most complex for its size of any here discussed.

Although at one point this area is only 1000 feet north of the Ganesha area, the two have almost nothing in common; there is a much greater variety of rock types in the Puddingstone area, yet no mappable unit was found surely common to them both; furthermore, although the Ganesha area is structurally relatively simple, the Puddingstone area has been folded and faulted on a fine scale so that outcrops of any one member are small and basement complex has been brought to the surface at several places.

The volcanics are described here in the sequence which best fits the known and inferred relations; the greatest uncertainty exists regarding the lowest members.

Early tuffs and basalts (not individually mapped).—The east-west ridge that culminates in summit 1011 about 1100 feet west of the dam is flanked by three outcrops of basement complex (including one of Mountain Meadows dacite). Among the vocanics in probable depositional contact with these the following three may be mentioned:

(1) Gray, inconspicuously vesicular basalt is exposed 500 feet east of summit 1011 in a breccia whose blocks display differing degrees of alteration. Microscopic examination reveals that the freshest samples are composed of 5–10 per cent plagioclase (An_{70}) in narrow ragged crystals few of which exceed 1 mm in length, 3–5 per cent bowlingite(?) in reddish pleochroic grains whose shapes suggest former olivine, and an intersertal groundmass of plagioclase, clinopyroxene, and glass. The vesicles are filled with chalcedony or opal or both.

(2) About 500 feet north of (1), or 800 feet northeast of summit 1011, is another (and possibly related) weathered dark-gray vesicular basalt breccia. The largest block seen, about 20 inches across, proved to be black glistening amygdaloidal olivine basalt in the interior, and because it is the most mafic rock known in this part of the Glendora volcanics a chemical analysis was made. The megascopic amygdules, chiefly dolomite with or without minor chalcedony(?), were removed insofar as possible from the analyzed sample, but the results (Table 2, S-615) indicate only partial success.

Petrographic Description. Volume per cent from micrometric analysis:

45.3	Plagioclase ($An_{60\pm}$) in laths less than 0.4 mm long
23.0	Augite in grains up to 0.15 mm across; may include some olivine
6.5	Magnetite in fine arborescent groups
4.6	Olivine phenocrysts, euhedral and subhedral, up to 1.5 mm, slightly altered
12.8	Glass, smokey
7.8	Other (alterations; some olivine)
100.0	

Intersertal texture with rare irregular vesicles lined with yellow altered glass(?) and sometimes filled with fibrous dolomite.

Chemical analysis: Table 2 (S-615).

(3) Near the basement complex along the north side of the ridge east of summit 1011, a yellow lapilli tuff under about 6 inches of soil is

composed of small subangular fragments of a fine-grained glassy volcanic whose plagioclase laths are calcic labradorite (An_{60-65}) and altered pumiceous material in an altered pasty matrix probably largely derived from original glass. No free plagioclase crystals were found.

Fine-grained biotite andesite (included in Tga on map).—A dark-gray, somewhat platy, and commonly brecciated very fine-grained andesite crops out in several small patches near Puddingstone Reservoir. In the Harwood Hills north and northeast of the reservoir it is apparently overlain by the more extensive hornblende andesite of that area, and along the crest of 1011 ridge immediately west of the dam it is at least locally in contact with basement complex. Fresh surfaces are light to dark gray; altered ones may be reddish, greenish, or bleached. With a hand lens, scattered minute feldspar laths are discernible on some specimens. Platy structure resulting from flow may be seen in some blocks but is not common. The best outcrops suggest origin as an autobreccia, the blocks of which are generally 1–2 inches across.

Petrographic Description. Fine biotite (hyalo) andesite. Percentages estimated:

40–70%	Plagioclase, An_{45-50} chiefly, but ranging from An_{40} to An_{60}; laths, average less than 0.3 mm long.
1–6	Biotite in shreds less than 0.03 mm long; some with normal pleochroism, some with Z′ = pale bluish green and X′ = pale brownish yellow; α and γ close to 1.548 and 1.595.
24–50	Groundmass glass with n below 1.54

Tridymite plates, a few over 1 mm, some probably primary, in many samples; γ' = 1.4775–1.4780, biref. less than 0.003

Felsitic rhyolite (included in Tgr on map).—At several places in the Harwood Hills, especially at their eastern end, there is exposed a finely laminated, gray to pink or red, felsitic rhyolite which is referred to as felsite to distinguish it from other banded glassy rhyolites. The individual laminae are a few millimeters or less thick and comprise nearly white layers alternating with dark-gray or (in pebbles) reddish-brown ones. In a few places a brick-red color is developed, and lamination is poor or absent. Fresh samples characteristically break with a clean conchoidal fracture. West of hill 1063 the felsite is usually brecciated and almost white, as for example in the exposures east and south of summit 1253. In the vicinity of summit 1009, west of the dam, there are many pockets of gray perlitic glass in breccias of blocks closely resembling the felsite and believed to be related to it. Locally, both east and west of the dam, blocks of very hard silicified fine breccia consist of angular fragments of gray felsite cemented with chalcedony.

Petrographic Description. Percentages estimated:

5–10%	Plagioclase (An_{18-25}) phenocrysts less than 0.15 mm across and abundant elongate microlites in parallel fluidal arrangement
90–95	Glass, n = 1.505–1.520
tr	Magnetite (detected with magnet on pulverized material)
tr	Tridymite plates, some twinned, lining or filling cavities; several percent of one specimen
tr	Chalcedony and cristobalite(?) filling minute cavities

Lighter layers of megascopic laminae are clearer glass with more numerous microlites; darker layers are clouded, probably somewhat devitrified.

One sample from junction of Walnut Ave. and Puddingstone Dr. has micropoikilitic texture with small patches of quartz similar to but smaller than those of hypersthene andesite of Ganesha area.

Most siliceous of rocks here studied.

Chemical analysis: Table 2 (S-496).

The 80+ feet of felsite exposed in hill 1063 is obviously a minimum thickness. A water well (No. 31 on map) drilled 1500 feet southeast of this hill bottomed in felsite at a depth of 405 feet, after passing through 403 feet of gravels and sands with minor clay. Because the fault at the north edge of the reservoir may continue eastward between this well and the hill, with upthrow on the north, this may not indicate a considerably greater thickness.

The felsite is probably older than the hornblende andesite described below, which places it in the lower half of the local succession.

Perlite (not separately mapped).—A distinctive bluish to lead-gray glass occurs in scattered outcrops in the Puddingstone area. West of the dam it usually occurs in splintery masses as pockets or lenses from less than an inch to more than 10 feet across in rocks somewhat resembling the felsite or intermediate

between the felsite and more glassy material. Perlite also occurs as scattered round particles up to half an inch across in some tuffs, notably the light-colored bedded tuffs exposed in the walls of Walnut Creek 0.5 to 1 mile downstream from Puddingstone Dam. A probable breccia pipe in Walnut Creek 0.5 mile below the dam contains many blocks of perlite and of banded glassy rhyolite with the same index of refraction but a higher proportion of aligned microlitic feldspar laths. The bands result largely from variation in the number of microlites, but the average proportion is somewhat less than in the felsite.

The chemical analysis given in Table 2 (S-597) is of a sample of perlite from 350 feet south-southeast of summit 1009 west of the dam. It came from a pocket about 1 foot across occurring, with many others of various sizes and shapes, in a breccia of banded glassy rhyolite blocks. Under the microscope it is seen to be composed of glass, with n close to 1.504, through which are sprinkled trichites, longulites, and rare plagioclase microlites, all totalling less than 2 per cent (estimated). Typical perlitic cracks are well developed (Pl. 2, fig. 4). In other samples the index of refraction ranges from 1.501 to 1.505. The high H_2O+ and perlitic structure are undoubtedly related; in other respects the composition is closely similar to that of the felsite. Thus megascopically, microscopically, and chemically the perlite seems to be a facies of the felsite.

None of the known occurrences is of clearly commercial proportions.

Hornblende andesite (included in Tga on map).—The most extensively exposed member in the Harwood Hills is a gray hornblende andesite porphyry and associated breccia. Probably not over half of the exposed part is massive.

The massive facies shows no obvious flow structure or systematic jointing. In over 95 per cent of the outcrops the hornblende phenocrysts have been altered and largely removed, leaving iron-oxide-lined cavities of tell-tale shape. The plagioclase phenocrysts commonly have chalky or dull centers and glassy rims, and the groundmass is cream-colored or yellowish. In the fresh rock the groundmass is lead gray, and the hornblendes are shiny and black save for slight marginal alteration around some crystals.

The breccia facies is somewhat similar to the hypersthene andesite breccia of the Mountain Meadows area, consisting of angular and slightly rounded blocks embedded in a matrix of similar material of fine sizes. The blocks are generally pale and altered; some are reddish or lavender-tinted in their outer portions (with notable oxidation of the hornblendes), and some of these are darker gray and somewhat fresher inside. The largest block seen was 35 inches long; the average is probably 2 inches or less.

The freshest sample seen was a block in the breccia exposed on the east side of a small ravine 650 feet N. 78° E. from summit 1253.

Petrographic Description. Micrometric analysis:

18.6%	Plagioclase phenocrysts, zoned, chiefly An_{52-55} but ranging from An_{62} (core) to An_{51} (rim); euhedral, up to 5 mm.
9.3	Hornblende, euhedral to subhedral, up to 3 mm, $Z \wedge c = 10°$, moderately pleochroic olive greens.
72.1	Groundmass, composed of (estimated):

Plagioclase	20–30%
Glass	10–20
Magnetite	1–3
Micropoikilitic spots	25–35
Tridymite and/or crystobalite	tr

In more-altered samples, feldspars commonly have thin sodic rims with extinction angles differing by more than 30° from the cores and the hornblendes are almost completely altered. In some, the hornblende is strongly pleochroic deep reddish-brown oxyhornblende with $Z \wedge c$ less than 5° and $\gamma - \alpha$ greater than 0.035.

Chemical analysis: Table 2 (S-519). Photomicrograph: Pl. 2, fig. 5.

The hornblende andesite is interpreted as resting on both the fine biotite andesite and the felsite. It is overlain by the tuff breccia with an irregular contact suggesting some relief on its upper surface before accumulation of the pyroclastics; this probably does not mean any great time interval, however, as there is no sign of weathering along the contact, and accounts of the eruptions of Mt. Pelee, Paricutin, and other volcanoes show that, in areas where the eruption of pyroclastics is interspersed with

that of flows, the surface of the latter is often irregular by construction.

The exposed hornblende andesite is at least 100 feet thick and may be over 500 (Pl. 6, section UV).

Tuff breccia (stippled Tgj on map).—The best exposures of the tuff breccia of the Harwood Hills are in road cuts along Cannon Avenue, 0.3 to 0.6 mile southwest of summit 1253. Here the rock is predominantly a greenish- or yellowish-gray unsorted mass of subangular blocks of intermediate and silicic volcanic rocks in an abundant tuff matrix.

The blocks range from less than an inch to 7 feet across, but very few exceed 12 inches. Blocks larger than half an inch across are estimated to constitute 35 to 50 per cent of the rock by volume. Some of the larger ones show pronounced flow banding. Porphyries of various colors predominate; chief among them (from hand-lens study) are a non-vesicular greenish-gray calcic(?) andesite and a vesicular andesite porphyry with irregularly shaped feldspar phenocrysts and very fine-grained dark groundmass. Doubtful felsite and tuff are rare constituents.

The matrix is composed of fairly well to poorly rounded small fragments of fine-grained and aphanitic volcanic rock, glass, pumice(?), and crystals of plagioclase (An_{55-60}), with some interstitial clay minerals. The color is light to dark yellowish gray. In one exposure the matrix is well bedded in thin laminae of fine lavender tuff with conchoidal fracture whose sorting and streaks of sand-size tuff indicate local reworking by water. This same feature has been observed in the matrix of the Johnstone Peak tuff breccia of the foothills.

In field mapping, which had to be based almost entirely on float, the tuff breccia was distinguished from the hornblende andesite breccia and volcanic conglomerate chiefly by the absence of hornblende andesite and felsite fragments, the abundance of calcic(?) andesite porphyry blocks, and, where possible, the tuffaceous matrix.

The origin of this tuff breccia poses some problems like those discussed in connection with the Johnstone Peak tuff breccia of the foothills. The chief differences seem to be that here a greater variety is represented in the blocks, and none of those seen have the radial cooling cracks so conspicuous in the foothills. In this example the possible origins include mud flow, avalanche, and peléan cloud.

Palagonitic tuff and pillow lava (Tgb with and without stippling on map).—About 0.3 square mile of lowlands west of the south end of Puddingstone Dam is underlain by weak brown palagonitic tuff and associated altered vesicular lavas with pillow structure.

The tuff, also exposed for a short distance east of the south end of the dam, is a rusty friable mass of residual glass and alteration products. A thin section was made of one sample (Pl. 2, fig. 6) that was harder than the rest because of the chalcedony lining its cavities. This piece is composed of fragments of highly vesicular glass and scattered vesicular rock fragments in a matrix of glass shards. The scattered euhedral and broken crystals of plagioclase are at least as calcic as An_{60} indicating that the material was probably basaltic. A sample from the south end of the dam lost 21 per cent of its weight when heated. Somewhat similar palagonitic tuff and tuff breccia was penetrated for about 1000 feet in the Los Angeles Brewing Co. Jones No. 1 well (Fig. 1, No. 12). Composed of about 90 per cent palagonitic material and 10 per cent labradorite crystals, the chemical analysis (Table 2) bears outs its derivation from largely basaltic glass.

Many exposures have a crude lamination which proved to give consistent attitudes in the form of a gentle anticline trending west-southwest from the south end of the dam.

Closely associated with the palagonite in the western part of the area is an altered rusty vesicular lava which was probably basalt; however, no samples fresh enough to establish this were found. Most outcrops show a more or less well-developed pillow structure; the rims are formed of granular altered glassy material, while toward the centers (especially in those 15 inches or more in diameter) there is some degree of radial jointing, and the vesicles become larger.

The palagonite and pillow lava are interbedded. They are overlain either by volcanic conglomerates or by shales and siltstones of the lower part of the Puente formation.

Volcanic conglomerates (Tgc on map).—

The volcanic conglomerates shown in the Puddingstone and Ganesha areas of Plate 1 are all predominantly andesitic. The largest clasts are generally less than 2 feet across, the average 1 to 4 inches. Most have rather flat faces with only slight rounding of the corners and edges; the smaller and softer ones show the highest degree of rounding and sphericity. Bedding, generally visible only in road cuts, varies from very crude to fairly good. The clasts are 50 to 90 per cent or more andesite; one of the commonest varieties is hornblende andesite either as greenish-gray boulders with fresh black hornblende needles or reddish-brown boulders with rusty pits of characteristic shape. A distinctive but not common clast is the laminated felsite. Other rock types include very fine biotite(?) andesite, very fine feldspar porphyry, and altered basalt(?). The matrix is everywhere highly tuffaceous, typically a fine lithic lapilli tuff. Conspicuously absent or rare are clasts of basement complex rocks and grains of quartz in the matrix. Closely associated with the patch 1200 feet south of summit 1233 south of Puddingstone Reservoir is a fine water-laid tuff containing locally abundant fish scales and termite coprolites. The latter, identified by Manley L. Natland, have the characteristic hexagonal shape (Stone, 1950); their occurence with fish scales is somewhat surprising.

The age and proper affinities of these conglomerates are still somewhat problematical. As now exposed none are clearly overlain by other volcanics. Their local derivation suggests but does not require an unconformity beneath them since they occur near the postulated shore line of the time; they are overlain, probably unconformably, by the Puente formation. Since the overlying Puente conglomerates are very different, with less than 10 per cent volcanic pebbles and abundant quartz sand in the matrix, it is believed that the volcanic conglomerates here described belong with the Glendora volcanics, in spite of the possibility that they may be younger than any of the primary deposits.

Area of complexly mixed volcanics and sediments.—Here, on a scale too small to show on Plate 1, volcanic rocks are intimately mixed with marine silstones and shales of lower Puente or slightly earlier age. Notable characteristics are the variety and high degree of alteration (including silicification) of most of the volcanics, the range in size of the slabs and pods of sediment and their twisted structure. It is estimated that about two-thirds to three-fourths of the volume is volcanics, and the remainder sediments. Some of the sediments are unusually rich in brown leaves and organic material. The whole suggests near-shore eruptions or submarine sliding or both.

Glendora Volcanics in Way Hill and Lone Hill

Way Hill is immediately north of San Dimas, and Lone Hill is a smaller prominence 1 mile west-southwest of it. Both are surrounded by Quaternary deposits.

Way Hill exposes a considerable variety of volcanics, the most abundant being tuff breccias and tuffs of various kinds with which are associated minor flows of fine-textured lavas. One of the larger bodies of tuff is exposed along Gladstone Avenue at the northwest end of the hill where it is in fault contact with fine-grained lavas and overlain by Quarternary terrace deposits. It consists essentially of lapilli-sized fragments of gray fine-grained andesite(?), pumice, and euhedral to subhedral crystals of andesine (An_{40-48}) in a yellow, almost waxy matrix containing somewhat altered glass with n near 1.470 and some unidentified grains. Rare biotite and possible minute fragments of basement complex rocks were also seen. The chemical analysis (Table 2, S-669) is consistent with the identification as a somewhat altered andesite vitric(?) tuff.

Midway along the south flank a reddish basaltic tuff stands vertically with northerly strikes; indistinct graded bedding suggests that the tops are toward the east. Along part of the north base of the hill there are outcrops of a fine-grained andesite similar to that in the foothills to the north in which the flow structure dips consistently northward. Three small patches of shale and sandstone around the base of the hill resemble those east of the Glendora South Hills and are tentatively assigned to the Topanga formation. No fossils were found in these, but a small outcrop of coarse sandstone in a road cut near the center of the hill has yielded a small molluscan fauna, and a sliver of

shale near the summit contains scales from 11 genera of fish. Both are interpreted as middle Miocene.

The structure of Way Hill is complex and uncertain—more like that of the Puddingstone area than any other. The representation of it shown on section LK is only a suggestion.

Lone Hill is composed chiefly of altered and silicified volcanic breccia.

Glendora Volcanics in the Glendora South Hills

To avoid ambiguity, the low ridge about a mile southeast of the town of Glendora labelled "South Hills" on the Glendora quadrangle is referred to as the Glendora South Hills (foreground, Pl. 5). The whole ridge is essentially sediments and interbedded volcanics of the middle Miocene Topanga formation. The oldest sediments are at the west end of the hills where several collections of upper middle Miocene Foraminifera and fish scales were made. The prevailing northeasterly dip (Pl. 6, section MN) exposes successively younger beds in that direction. No good fossils were found in the eastern half of the hills, but, in the absence of any clean-cut change which might be mapped as the base of the Puente formation, all the exposed beds were included in the Topanga.

A small area (less than 0.1 square mile) at the west end of the hills is underlain by altered hornblende andesite flows and breccia (Tga on map). The rock is a light-gray porphyry in which 1- to 2-mm phenocyrsts of feldspar are the most conspicuous constituents. Narrow decomposed hornblende crystals up to 7 mm long but averaging 3 mm are less abundant and usually identifiable only by their shapes. In many cases the alteration, which is evidently largely deuteric rather than related to modern weathering, has given the rock a characteristic pinkish hue. Under the microscope the feldspar phenocrysts are usually zoned; the more calcic centers are noticeably more altered than the rims. Composition ranges from about An_{30} to An_{50}, and the greatest volume is calcic andesine. In one sample studied the hornblende remnants were oxyhornblende.

This rock probably flowed into the sea, as shown by some details of its structure as well as its interbedding with marine sediments. In a prospect pit at the southwest end of the hills it is traversed by irregular thin stringers of brown micaceous siltstone interpreted as mud that was squeezed up into the brecciated flow as it advanced. More striking, however, are the many sandstone dikes which cut the andesite. They range from an inch to about 3 feet wide and are all steeply dipping or nearly vertical. One, traced for over 600 feet with a uniform width of 4 to 8 inches, is an arkose composed of quartz, oligoclase, biotite, muscovite, hornblende, garnet, opaque minerals, nonvolcanic rock fragments, and other grains, all quite fresh, in a sparse opal cement which makes it so resistant that it stands up as a low wall along part of its outcrop. Based on mineral suites of the available sands, the fracture might have been filled either from above or below.

In some places this andesite is underlain by a calcite-cemented lapilli tuff composed predominantly of glass with index of refraction close to 1.59 and containing bytownite; maximum thickness is about 30 feet. The hornblende andesite is about 200 to 300 feet thick (section MN) and may be the same as the one in the Harwood Hills except for having been erupted into the sea.

The south-central portion of the hills (Tg on map) is chiefly fine-grained flows and autobreccias close to the boundary between basalt and andesite with which are found some basaltic(?) and andesitic tuffs whose well-bedded character in some outcrops further indicates subaqueous accumulation.

Of the two smaller areas of volcanic rocks at the east end of the hills, the southern one is composed of fine-grained flows similar to those just described and the northern is a complex association of coarse basalt breccia, dacite(?), and highly altered fine-grained volcanics and sediments. The relations of these indicated on section MN is little better than a guess. Future study and/or excavations may uncover evidence that this was a small vent or an area of secondary activity.

Glendora Volcanics in the Foothills Area

General statement.—This area includes the volcanics exposed in the San Gabriel Mountains

foothills, north of U. S. Highway 66 (Pl. 1). They extend from near the mouth of San Gabriel Canyon on the west to just beyond San Dimas canyon on the east with an aggregate outcrop area of approximately 3 square miles. The largest portion, between Dalton and San Dimas canyons, is illustrated in Plate 5.

The Foothills area differs from those already described in the greater exposed thickness of the series as a whole and the larger volume of andesitic flows and tuff breccias. Although there are some places, such as along the line of section LK, where the sequence is in regular order for 1000 and perhaps 2000 feet stratigraphically, there are many others where the detailed structure is quite complicated. Many of the rocks were evidently originally small irregular tongues in variable sequence; extensive ancient and modern landsliding have further complicated their distribution. No fault is shown on Plate 1 for which there was not direct evidence; none were used to account for abrupt changes just because that might be the habit of the area. Accordingly many are shown as very short, and unquestionably many were missed because of their inconspicuousness within rocks of broken and irregular lithologic character.

A single reliable volcanic sequence for all of this area probably does not exist; however the following generalizations seem to hold:

The fine-grained andesite is the oldest volcanic rock wherever it occurs. Where its base is exposed it rests only on basement complex. However, other members may also rest on the basement.

Only two episodes of basalt eruption are known. One came early in the sequence and is exposed south of Johnstone Peak. The other is probably late as it is interbedded with upper middle Miocene sandstones and shales 0.6 mile northeast of Azusa. The age of several intrusive basaltic bodies is unknown.

The calcic andesite flows and tuff breccias recur several times in the sequence and were probably erupted in alternation.

Although the general structure is a homocline dipping 30° or more southward, the individual members of the sequence are not continuous over large areas, probably because of intermittent eruption onto rough terrain and/or contemporaneous erosion and/or the presence of several active vents.

The members are described in the inferred sequence of first eruptions for each rock type; most of them recurred two or more times.

Fine-grained andesite (Tgf on map).—In the foothills area the oldest exposed member of the Glendora volcanics is a fine-grained to very fine-grained massive andesite and associated red tuffaceous breccia. It rests on basement complex wherever its base was seen and is not known to recur higher in the sequence in this area. At different places it is overlain by the Johnstone Peak tuff breccia, basaltic lava, or calcic andesite flows. Its outcrop thickness ranges from less than 100 to more than 200 feet; the probable maximum is 500 to 800 feet.

The fine-grained andesite is usually light to dark gray on fresh surfaces and pale yellowish gray to greenish gray where weathered. Flow structure is almost always present, indicated by platy parting, corrugations on weathered surfaces, or the alinement of minute feldspar laths or collapsed vesicles. Much of the rock is so fine-grained that the small feldspar phenocrysts are the only minerals identifiable with a hand lens. Close jointing facilitates disintegration so that smooth slopes, without ledges or large float blocks, characterize areas underlain by this member. In the field it is one of the most readily recognized of all the Glendora volcanics.

Petrographic Description. Estimated average composition, with ranges in parentheses:

71%	(60–85)	Plagioclase An_{44} (An_{15}–An_{55}), phenocrysts (up to 0.5 mm) in general more calcic than groundmass; zoning much less common than in calcic andesite described below
15	(10–25)	Glass, n below 1.54
6	(1–10)	Augite in minute grains
5	(3–8)	Opaques, at least partly magnetite
3	(1–5)	Chlorite, altered biotite, etc.

Texture: rare phenocrysts in glass with minute plagioclase laths in fluidal arrangement.

Chemical analysis: Table 2 (S-339). Photomicrograph: Pl. 3, fig. 1.

Near the mouth of Big Dalton Canyon, and especially east of it, a considerable quantity of orange-red to brownish-black tuffaceous volcanic breccia appears to be more closely related

to the fine-grained andesite than to any other member of the volcanic series. Most outcrops are weak and crumbly; a few are extremely hard owing to cementation by chalcedony and other secondary minerals. The rock is made up chiefly of small, flat, angular or slightly rounded slabs and fragments ½ to 2 inches across. The streaky, reddish, aphanitic matrix is largely grains of andesine (An_{40-50}). The fragments are mostly fine-grained andesite, but locally include some coarser-textured calcic andesite. Rarely, in close proximity to basement complex, the breccia includes pebbles and cobbles of aplite and other basement rocks, ranging in size down to individual grains of quartz. Locally it shows bedding in the finer pyroclastic parts. This red breccia is interpreted, without much confidence, as an early tuffaceous mud flow. On Plate 1 it is not distinguished from the massive fine-grained andesite flows.

Calcic andesite flows (included in Tga on map).—In the mapped area calcic andesite is one of the most abundant rock types in the Glendora volcanics. It is gray, often with a lavender tint, and studded with abundant feldspar phenocrysts. It occurs in units at least as much as 200 feet thick and possibly much more, and recurs three or four times in the sequence. It lies on fine-grained andesite, tuff breccia, and basaltic lava at different places in the foothills and is overlain by tuff breccia almost exclusively.

The typical hand specimen consists of abundant, well-formed, white plagioclase phenocrysts about 1 mm long in a dark-gray or brownish-gray aphanitic groundmass. In some areas altered hypersthene phenocrysts 2 mm long are fairly abundant, sometimes in radiating groups. Vesicles are rare and small. In the most massive outcrops the rock is usually traversed by well-developed joint systems. Flow banding can locally be seen.

Petrographic Description. Estimated average composition, with ranges in parentheses:

28% (11–45) Plagioclase phenocrysts An_{58} (An_{25}–An_{68} in different zones and different specimens); up to 2 mm

43 (17–60) Groundmass plagioclase An_{45} (An_{21}–An_{57} in different zones and different specimens)

8 (3–12) Hypersthene phenocrysts, altered in most specimens

4 (1–10) Augite, mostly in groundmass

6 (2–10) Opaques

11 (0–15) Glass, n below 1.54

Chemical analysis: Table 2 ($S\text{-}333_2$). Photomicrograph: Pl. 3, fig. 2.

A rusty-weathering massive breccia of calcic andesite fragments, some of which are markedly vesicular, lodged in a matrix of similar but finer-grained material, is locally associated with the calcic andesite. It is interpreted as an autobrecciated facies of this member, probably representing the margins of a flow, and is not distinguished on the map.

Johnstone Peak tuff breccia (included in Tgj on map).—The accumulations of tuff that occur at various horizons in the Glendora volcanics of the foothills contain numerous scattered subangular fragments and blocks ranging in diameter from a fraction of an inch to over 10 feet. The proportion of blocks to tuff matrix ranges from about 10 to over 50 per cent with the average estimated to be 15 to 20 per cent (Pl. 4, fig. 1). Following the suggestion of Norton (1917), adopted by Wentworth and Williams (1932), these deposits are termed tuff breccias and are here named for Johnstone Peak, south of which (Pl. 1) are the most typical exposures. They lie between and upon the three lava members of the volcanic series and range in thickness from about 100 to 700 feet (Pl. 6, section JK). Their poor consolidation results in weak topographic expression, few outcrops, and susceptibility to landsliding.

In most outcrops the deposit is chaotic with no sign of bedding, but locally lenses and thin streaks of bedded tuff matrix indicate reworking of the fines.

About half the tuff matrix (Pl. 3, fig. 3) is labradorite (An_{55-60}) in fragments many of which show crystal faces; about 15 per cent is altered glass; and the remainder consists of green pyroxene, rather rare, black, slaggy-looking grains of some manganese-bearing mineral, and unidentified particles. No inclusions of basement rocks or grains of quartz were seen. A few samples of the matrix were screened; everything passed a 2-mm screen, and roughly equal portions were retained on the 1-mm, 0.5-mm, and 0.1476-mm screens, with a small amount finer than 0.1476 mm.

The average size of the blocks is probably

well under 1 foot. Most are fresh, nonvesicular calcic andesite with hypersthene (Pl. 3, fig. 4), but a few greatly altered blocks may have had more mafic compositions. Occasional blocks show vesicular structure, commonly with the vesicles partially or completely filled with chalcedony or opal. Many of the larger blocks (over 3 feet long) tend to be tabular or bun-shaped, with their flat bottoms often parallel to the bedding in the tuff.

The various possible modes of origin for deposits of this type have been systematically discussed by Anderson (1933, p. 245–261) who groups them according to whether or not water played an essential role in their transportation. In most of the Johnstone Peak tuff breccia the matrix is clean, pure, and fairly fresh without the accidental inclusions, oxidized zones, or muddy streaks and crude bedding that to some degree characterize volcanic mud flows. Whereas parts of it may have been subsequently redistributed as mud flows, it is believed that this was not the dominant original mode of accumulation. In this connection the following considerations are pertinent:

The high proportion of tuff matrix in the Johnstone Peak tuff breccia suggests either the avalanching of an accumulation of interbedded tuffs and flows on the steep side of a volcanic cone or a peléan eruption like the *nuées ardentes* of Lacroix (1904). The critical feature of the latter is the high temperature attending their movement during which evolution of gases is believed to lend remarkable mobility. Charred and carbonized wood has been used as evidence of this heat in some cases, but such were not found in the Johnstone Peak area.

However, a notable feature of the larger exposures of the Johnstone Peak tuff breccia is the presence in it of radially jointed blocks (Pl. 4, fig. 2) of hypersthene andesite ranging from 8 inches to 10 feet across. The frequency of occurrence of such blocks can be approximated by dividing the total area of outcrops of this member (practically limited to vertical road cuts and a few steep canyons) by the number of blocks seen. The result is one for every exposure approximately 40 by 100 feet, which is considered a minimum because of the combination of favorable circumstances necessary; there must be many more that were too coherent to allow a part to fall away and reveal their inner structure as they were undermined by erosion.

These joints must have developed after the blocks reached their present resting places, for the chaotic nature of the accumulation and the size of the blocks require that the transportation must have been rather violent. Yet the blocks show no sign of alteration along the

PLATE 2.—PHOTOMICROGRAPHS OF GLENDORA VOLCANICS SOUTH OF THE FOOTHILLS AREA (All ×17)

FIGURE 1.—Glassy block from Elephant Hill breccia. S-548-2. Ordinary light.

FIGURE 2.—Hypersthene andesite, Ganesha area. S-468. Phenocrysts of plagioclase and hypersthene; micropoikilitic groundmass. Ordinary light.

FIGURE 3.—Fine-grained basalt porphyry, Ganesha area. S-481. Phenocrysts of augite, hypersthene, and plagioclase. Ordinary light.

FIGURE 4.—Perlite, west of Puddingstone dam. S-453. Ordinary light.

FIGURE 5.—Hornblende andesite porphyry, northeastern Harwood Hills. S-519. Ordinary light.

FIGURE 6.—Palogonitic tuff, west of Puddingstone dam. S-447. Ordinary light.

PLATE 3.—PHOTOMICROGRAPHS OF GLENDORA VOLCANICS FROM THE FOOTHILLS AREA (All ×17)

FIGURE 1.—Fine-grained andesite, Mull Canyon, foothills area. S-339-1. Shows fluidal texture and comparatively rare plagioclase phenocryst. Ordinary light.

FIGURE 2.—Calcic andesite, Wildwood Canyon, foothills area. S-333-2. Crossed nicols.

FIGURE 3.—Matrix of Johnstone Peak tuff breccia. S-438. Dark area is portion of an andesite fragment; small circles are bubbles in mounting medium. Ordinary light.

FIGURE 4.—Hypersthene andesite block in Johnstone Peak tuff breccia. S-345. Ordinary light.

FIGURE 5.—Basalt lava from upper Sycamore Canyon. S-347. Crossed nicols.

FIGURE 6.—Basalt dike(?) in Topanga(?) sandstones northeast of Azusa. S-422. Ophitic texture of gray pyroxenes and light feldspar laths; large white grain is olivine. Ordinary light.

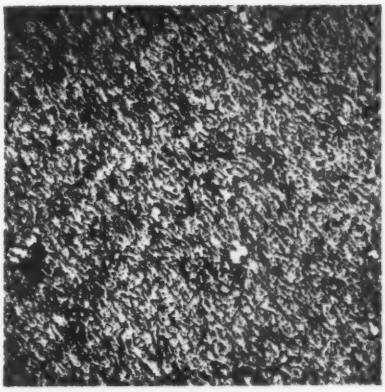


FIGURE 1

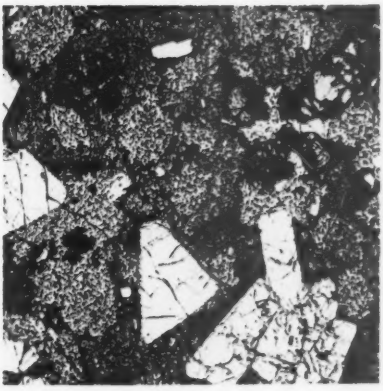


FIGURE 2

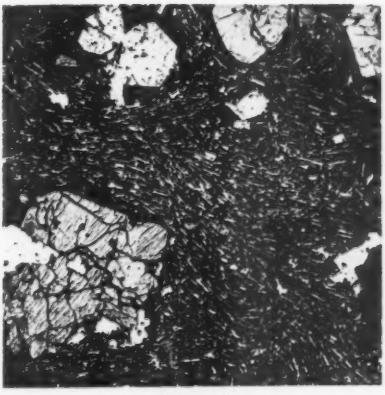


FIGURE 3

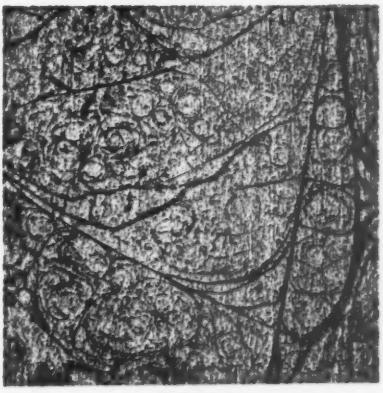


FIGURE 4

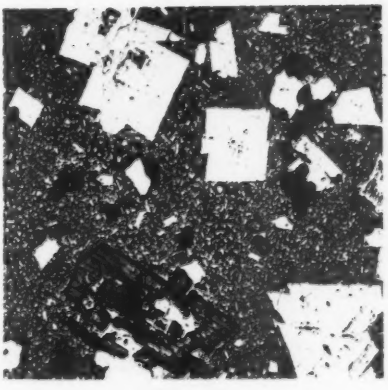


FIGURE 5

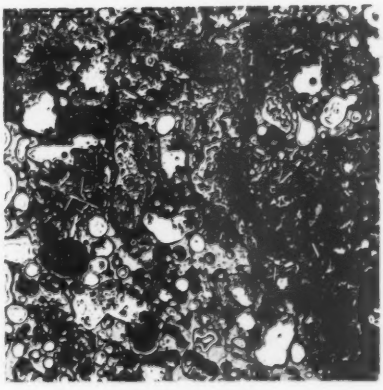


FIGURE 6

PHOTOMICROGRAPHS OF GLENDORA VOLCANICS SOUTH OF THE FOOTHILLS AREA

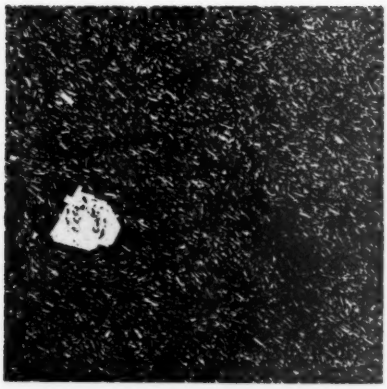


FIGURE 1

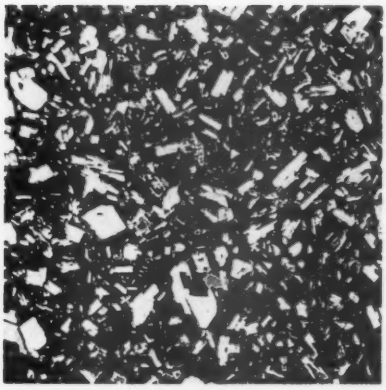


FIGURE 2

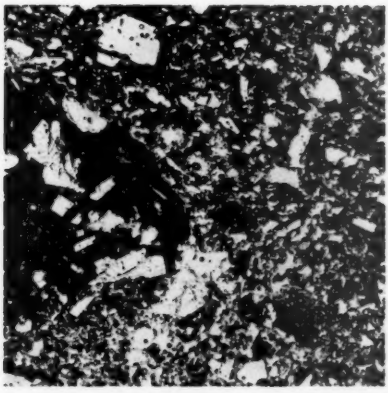


FIGURE 3

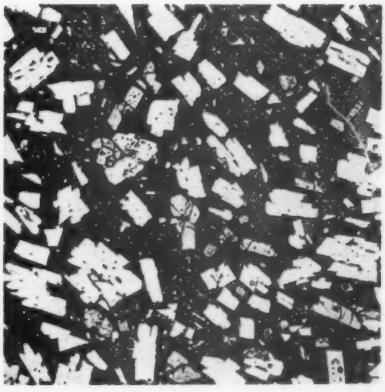


FIGURE 4

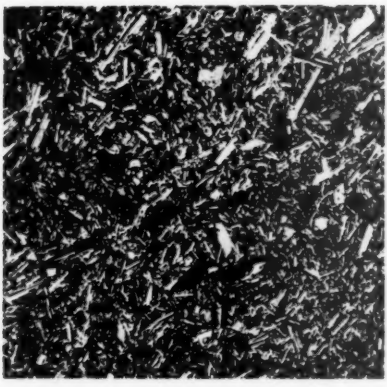


FIGURE 5

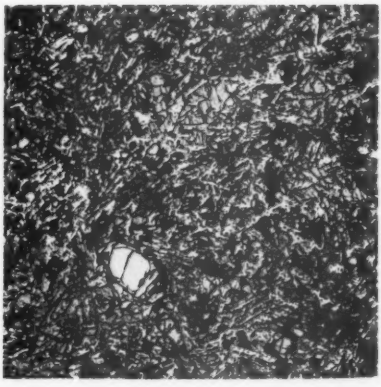


FIGURE 6

PHOTOMICROGRAPHS OF GLENDORA VOLCANICS FROM THE FOOTHILLS AREA

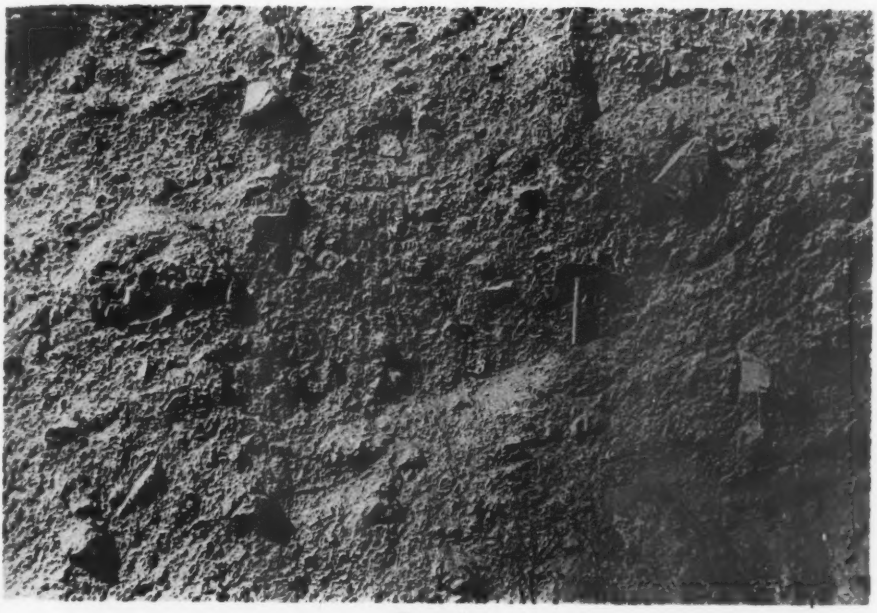


FIGURE 1

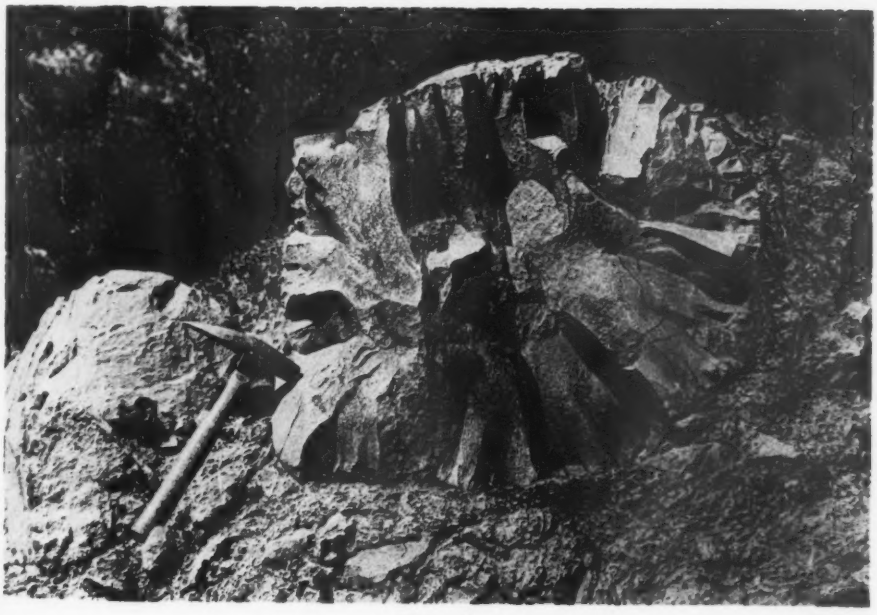


FIGURE 2

JOHNSTONE PEAK TUFF BRECCIA

AERIAL VIEW OF GLENDORA SOUTH HILLS AND FOOTHILLS OF SAN GABRIEL MOUNTAINS

joints; the hypersthenes of the outermost rim may be largely altered, but this is believed to have been deuteric since in no case is it known to be related to weathered surfaces.

It is believed that these joints are the result of contraction upon cooling in place, and accordingly that the blocks were hot when emplaced. Recent examples may be seen in and about Lassen Peak crater where blocks of the 1914 dacite from 1 to 20 feet across exhibit similar joints; change in temperature seems to be the most likely mechanism here.

The view that these temperatures obtained throughout the mass of tuff breccia rather than that the hot blocks were tossed out at intervals onto an accumulating blanket of ash is supported by the chaotic character of the deposit as a whole, the large size (up to 15 feet) of some of the blocks, and the observations of some who have witnessed or studied historic peléan eruptions. Thus Lacroix (1904, p. 380, Fig. 161) has described and figured large blocks with cooling cracks in the type area of the *nuée ardente*, and Perret (1935, Fig. 24) has photographed them at the same place. MacGregor (1938, Pl. 5, fig. 32) in his account of Montserrat reported large blocks with radial cooling cracks "... embedded in *nuée ardente* agglomerate of the Soufrière Hills."

Locally, in zones from a few inches to a few feet thick, the tuff breccia is free of blocks and well bedded. These spots, which are rare and in conspicuous contrast with the rest of the deposit, are interpreted as the result of local contemporaneous reworking of the tuff breccia in which the fines have been winnowed out and redeposited by the runoff from rainstorms.

The Johnstone Peak tuff breccia probably represents the eroded remnants of a deposit formed by several eruptions of *nuée ardente* type between some of which there were local rainstorms. Although some eruptions have been credited with inducing rainfall, Neumann (1934a, p. 173; 1934b, p. 90) has emphasized that sometimes they seem to prevent it, so no implication of causal relation between the eruptions and the rains is here intended.

The analyzed sample of a typical block (Table 2, S-345) was taken from the center of one like that shown in Plate 4, figure 2; its texture is illustrated (Pl. 3, fig. 4). The matrix analyzed (Table 2, S-438) was collected about 100 feet away.

The tuff breccia recurs several times in the sequence; probably no single occurrence exceeds 400 to 500 feet in thickness.

Basaltic lava (included in Tgb on map).—A distinctive dark basaltic lava is exposed between Morgan and Ham canyons where it occurs as a single mass composed of several flows with a maximum aggregate thickness of about 150 feet. It rests on basement complex, fine-grained andesite, or tuff breccia and is overlain by tuff breccia or calcic andesite. Near the Rainbow Club a mile northeast of Azusa a layer of vesicular basalt 100 feet thick occurs in sandstone of the Topanga formation from which upper middle Miocene Foraminifera have been reported.

These basalts are usually black and vesicular, generally weathering dark purplish brown or dull gray. Fresh surfaces commonly show a few 0.5- to 1-mm phenocrysts of feldspar and pyroxene, the latter usually rusty, in a dark groundmass. Olivine was found in one sample. Vesicles, where present, range from 0.5 to 12 mm in diameter and may be open, lined with opal, or filled with opal, chalcedony, or calcite. In a few places there is some fragmental material associated with the flows.

PLATE 4.—JOHNSTONE PEAK TUFF BRECCIA

FIGURE 1.—Johnstone Peak tuff breccia. A typical exposure in road cut south of Johnstone Peak and west of Ham Canyon.

FIGURE 2.—Andesite block in Johnstone Peak tuff breccia. Well developed radial joints interpreted as cooling cracks. Steep ravine south of summit 2428, south of Johnstone Peak.

PLATE 5.—AERIAL VIEW OF GLENDORA SOUTH HILLS (FOREGROUND) AND FOOTHILLS OF SAN GABRIEL MOUNTAINS

Big Dalton Canyon on the left and San Dimas Canyon at the right. Johnstone Peak is the middle distance summit with the lookout tower between these canyons. Taken from a point about 2 miles south of Glendora, looking northeastward.

Petrographic Description. Estimated average composition, with ranges in parentheses:

67%	(55–75)	Plagioclase, weighted average An_{63} (An_4–An_{74}), zoned, with sodic rims; mostly in groundmass
11	(5–15)	Augite, mostly in groundmass
5		Hypersthene pseudomorphs (in 3 sections), wholly altered to iron oxide etc.
1		Olivine, ragged phenocrysts (7% of one section)
8	(5–15)	Opaques, chiefly magnetite
8		Chlorite, opal, chalcedony, etc., exclusive of vesicle fillings

Holocrystalline, with moderately well developed flow structure.

Chemical analysis: Table 2 (S-347). Photomicrograph: Pl. 3, fig. 5.

Hornblende andesite (tuff) breccia (included with Tgj on map).—This member is known chiefly from two small areas south of Johnstone Peak where it closely resembles the Johnstone Peak tuff breccia in general color, unsorted character, and the presence of some blocks with radial cooling cracks. One of these, 10 feet across, was the largest seen in either member. This member is distinguished by the slightly lower proportion of tuff matrix to blocks and the fact that half or more of the latter contain abundant shiny black hornblende needles 3 to 6 mm long. In some of the blocks the interior is fresh, but in the outer portion the hornblendes have been altered and the groundmass has acquired a characteristic dark reddish hue. Because this may be true of one block and not of another close beside it, it is believed to be the result of alteration before emplacement—perhaps in blocks that fell back into the crater instead of on the outer slopes of the vent.

The blocks consist of about 15 per cent subhedral to euhedral phenocrysts of zoned labradorite (An_{50}-An_{60}), and 5 per cent hornblende in a fine-grained groundmass of feldspar laths, glass, magnetite(?), and unidentified birefringent specks.

The deposit may represent the remains of a late variant of the Johnstone Peak tuff breccia. On the south slope of summit 2989, 3000 feet southwest of Johnstone Peak, it is locally exposed as an accumulation 30 to 40 feet thick between Johnstone Peak tuff breccia below and calcic andesite flows above (Pl. 6, section JK). Another area of it is next to the Puente shales west of San Dimas Foothill Park.

Dikes and other possible intrusive masses (Tgd and Tg on map).—Doleritic and diabasic dikes cut the calcic andesite, basaltic lava, and Johnstone Peak tuff breccia as well as basement complex. Those cutting the calcic andesite are often amygdaloidal and highly altered; they are not differentiated from the calcic andesite on Plate 1.

Those cutting the tuff breccia are as weak topographically as their host but in many places are marked by indurated contact zones which stand up as low walls. Such a one is exposed on the southeast slope of summit 2989, south of Johnstone Peak; the microscope shows that it has the texture of a dolerite or coarse basalt and is about 70 per cent plagioclase (An_{55}), 20 per cent augite, 10 per cent magnetite and other minerals.

Dikes were seen cutting the basalt only in lower Sycamore Canyon where they are deeply altered and involved in complex structure. They were not mapped or examined under the microscope.

The dikes in the basement complex between San Dimas and Liveoak canyons are all dark and fine-grained. One, where it crosses a small canyon, shows fairly well developed columnar structure. Three samples of these rocks studied under the microscope were all basalts close to andesite (feldspars An_{50} to An_{60}).

One mile east of the mouth of San Gabriel Canyon and 150 feet north of Sierra Madre Avenue there are abundant large float blocks of fresh hard basalt confined to a narrow east-west zone about 400 feet long. Because it is the most uniformly fresh basalt known in the region, and despite the uncertainty as to whether it is a dike, sill, or flow and the possibility that it may be of post-Topanga age, it was submitted for analysis (Table 1, S-422).

Petrographic Description. Micrometric analysis:

46.2%	Plagioclase An_{65}	
19.4	Clinopyroxene, up to 2 mm across	
14.0	Olivine	
20.4	Other, estimated as:	
	14.0	Alteration products
	4.6	Fine groundmass
	1.5	Magnetite
	0.3	Isotropic

Texture: ophitic

Chemical analysis: Table 2 (S-422). Photomicrograph: Pl. 3, fig. 6.

Two thousand feet southeast of Johnstone Peak is a 500-foot patch of light-colored biotite dacite breccia. Maximum size of individual blocks is at least 4 feet. These are somewhat altered and are composed of about 80 per cent oligoclase (An_{20}-An_{30}) in euhedral and broken crystals and 5 to 10 per cent quartz in rounded grains, the remainder being low-index glass, biotite, magnetite, and alteration products. With the possible exception of one sample from a water well near the mouth of Ham Canyon, such a rock is unknown in the extrusive Glendora volcanics. Three dacite dikes cutting basement complex within 1500 feet of it are similar and it is not unlike the Mountain Meadows dacite except in texture. In two places, where it is crossed by a small ravine, it seems to have fairly steep contacts. (Outcrops are poor.) It may be a pre-Glendora volcanics vent (perhaps related to the Mountain Meadows dacite) which was topographically high when the first Glendora volcanics were erupted in its vicinity.

Near its mouth Big Dalton Canyon is joined from the north by Mystic Canyon. Just above the junction the latter cuts through a 600-foot patch of dark volcanics surrounded by basement complex. The exposures are good enough to establish that the contacts are steep on several sides and probably so all around. This rough cylinder is filled with a nearly black mass of hard tuff or fine breccia. Accidental fragments of basement complex occur throughout, ranging from abundant pieces several inches across near the contacts to rare grains of quartz or clusters of grains of quartz and feldspar near the center. At one place where the contact is well exposed there is a crude gradation from massive basement complex in which the gneissic structure and feldspathic dikes are undisturbed, through a broken zone 6 to 12 feet wide in which volcanic and basement rocks are thoroughly mixed in angular and subangular fragments averaging less than 3 inches (but reaching 2 feet) across, into the fine breccia or tuff of the main mass in which basement inclusions are a minor constituent. A tongue of fine augite-biotite basalt (or calcic andesite) porphyry in this mixed-rock zone looks like an aborted dike. Elsewhere, as near the southwest contact, the cylinder is partly filled with fine-grained basalt or andesite composed of approximately 76 per cent plagioclase (An_{50}), 15 per cent biotite, 5 per cent glass, 4 per cent opaque minerals, and rare pyroxene(?). The texture shows extreme parallelism of the plagioclase laths traversed by narrow zones in which they lie oblique to the general trend, probably as a result of viscous shear.

This is believed to be the best-established of the possible vent fillings so far discovered. If so, it seems most likely that the last rock erupted was biotite andesite or basalt; the nearest rock of this type known today is in the Puddingstone area, 5 miles away.

Inter-Area Correlation of the Glendora Volcanics

None of the major members of the exposed Glendora volcanics—*i.e.*, those which are most widespread in their respective areas and might be expected to outcrop in others—is exactly duplicated outside its "type" area. On the basis of lithologic similarity and some rather vague ideas on sequence, 5 of the over 20 volcanic types described may be compared as follows:

Fine-grained biotite andesite occurs in the Puddingstone area and in one small outcrop at the east end of the Spadra area. Both probably lie on basement complex. If they were once joined, no evidence has been found of this in the intervening Ganesha area where it might be expected at the horizon of the pre-hypersthene andesite tuffs.

The fine-grained andesite (without biotite) from the north edge of Way Hill and possibly that from the east end of the Glendora South Hills may be related to the widespread fine-grained andesite of the foothills area. The chief objection to this is that the first is probably and the second surely a late member of the local sequence, whereas the last is clearly at the bottom of its pile. However, correlation of the first two is a possibility.

Hornblende andesite occurs in the Harwood Hills as flows and breccia, in the Texas Company's Covina 27–1 well (No. 8 on Pl. 1) as

breccia at 1150 feet (one core only), as probable submarine flows in the western Glendora South Hills, and in the San Gabriel foothills as tuff breccia. Petrographically the four are closely similar, except that the foothills variety occurs in a highly tuffaceous matrix. Each occurs near or above the middle of its local sequence. It seems probable that the Covina 27–1, Glendora South Hills, and Harwood Hills samples represent the same unit. If so, several hundred feet of Topanga sandstone apparently laps out eastward between the first two and the last-named localities.

A yellowish altered lapilli tuff in the pre-hypersthene andesite deposits of the Ganesha area is similar to one occurring west of Puddingstone Dam and possibly to another at the west end of Way Hill. The first two are close to or on basement complex; the last is faulted against tuff breccia and of unknown relative age. The first two, and possibly all three, may be remnants of an early tuff that blanketed much of the region.

Tuff breccia, *sensu lato*, is one of the most common rock types in the region. It occurs in all save the Spadra area, although the amount in the Glendora South Hills is very small (one 30- to 40-foot zone in the central part). All occurrences are dominantly andesitic, but there are differences of detail such as maximum size of blocks, minor rock types present, degree of sorting, and proportion of matrix to blocks. Only in the higher parts of the foothills area are blocks with radial cooling cracks known. Although the recurrent eruptions demonstrated for the foothills area weaken its stratigraphic significance, it is noteworthy as the most widespread of the ejectamenta exposed today. Some of the cores from the Covina 27–1 well are probably tuff breccia, and most geologists who have dealt with the subsurface volcanics in the Los Angeles Basin believe far more of them are fragmental than massive; the writer's observations substantiate this.

Petrology and Chemical Composition of the Glendora Volcanics

The Glendora volcanics are extensively altered, but the principal changes are not due to weathering even though weathering has been favored by the texture of the tuffs and the considerable fracturing of the lavas. Almost without exception hot gases and solutions active during and perhaps immediately after eruption are believed to have caused the greatest changes. In the Johnstone Peak tuff breccia, for example, many blocks possess a sharply defined rind about half an inch thick which is lighter-colored than the interior and apparently arises from slight devitrification of a small amount of interstitial glass in the groundmass. Blocks exhibiting this feature occur beside others lacking it.

Further, the hypersthene in the calcic andesite flows and several tuffs implies similar deuteric alteration. Deep road cuts and quarries yield samples with just as poorly preserved hypersthene as those from near the surface; altered hypersthene commonly occurs in lavas whose feldspars are fresh. In fact, fresh crystals of hypersthene are more abundant in most tuffs than in most lavas; in the Ganesha area they are especially abundant in some of the yellow altered-looking pre-hypersthene andesite tuffs resting on basement complex where Recent weathering has probably been quite effective. The explanation probably lies in the escape from the tuffs of volatiles which, better retained in the lavas, have there destroyed most of the hypersthene. Fuller (1938) explained in similar fashion joint blocks of basalt whose olivine was fresh near the surface and decomposed in the interior.

Direct evidence of mineralization exists in several areas, notably silicification in the Puddingstone and Spadra areas and introduction of sulphides locally in the Foothills area and generally in the Spadra area.

Throughout this work it was assumed that in a given rock the larger the percentage of glass the more calcic will be the plagioclase phenocrysts (or their outermost zones). Thus several rocks bearing labradorite phenocrysts were classed as andesites. The chemical analyses corroborate this.

In Figure 3, 14 analyses of members of the Glendora volcanics and one of Mountain Meadows dacite (Woodford *et al.*, 1946, p. 527) are plotted on a Niggli-Becke-Bacon diagram (Bacon, 1947). Also shown, as guides to common usage, are points representing Tröger's

(1935) type rhyolite, dacite, andesite, and basalt, and Daly's (1933) averages for these same rocks. The comparisons thus rendered graphic show that, in general, the Glendora volcanics, while of normal silica content, are notably low in potash—a feature already known to be characteristic of the Mesozoic plutons of this province (Larsen, 1948). Further, they are relatively poor in ferro-magnesian constituents for the composition of their plagioclases; *i.e.*, they are somewhat leucocratic or feldspathic. Both of these features are reflected in the fact that on a "variation diagram" the curves for $Na_2O + K_2O$ and for CaO cross at about 65 per cent SiO_2, placing the volcanics in the calcic series (Williams *et al.*, 1954).

Most of the conspicuous variations in the individual analyses can likewise be explained petrologically. The tuffs (S-669, S-438, and Jones 1) stand apart in having high ferric and low ferrous iron, as well as low CaO and alkalies and high water. The perlite has nearly six times as much H_2O+ as the felsite to which it is believed to be related. The high CO_2 of the basalt from west of the dam (S-615) reflects carbonate amygdules.

In general, considering the volumes of the various members as now exposed, the analyses show that the Glendora volcanics are overwhelmingly andesitic. Two of the four basalts (S-422 and S-615) are rare, and a third (S-347) is minor, two of the three rhyolites are rare and a third (S-347) is minor, two of the three rhyolites are rare (S-597 and S-673), and the third minor (S-496), while six of the seven andesites (all except S-669) are major units. (The older Mountain Meadows dacite is also a major rock unit.)

Topanga Formation

Rocks assigned to the middle Miocene Topanga formation on faunal and lithologic bases (Hoots, 1931) are best exposed in the Glendora South Hills (Pl. 6, section MN), in the vicinity of the Rainbow Club northeast of Azusa (Pl. 6, section FG), and around Buzzard Peak in the central San Jose Hills.

In the Glendora South Hills the Topanga sediments are about 2000 feet thick, of which an estimated 35–40 per cent is interbedded volcanics. The sediments consist predominantly of massive and bedded sandstones. Lenses of conglomerate and isolated boulders of basement and volcanic rocks as much as 6 feet across are common, especially in the western, and to a lesser extent in the eastern part of the hills. The largest boulders do not occur in the conglomerates but rather in scattered groups embedded in massive sandstone with or without a few associated pebbles less than an inch in diameter. The mid-portion of the Hills includes more fine sediments of silt size. Calcareous cements are characteristic. Carbonaceous and silicified wood fragments are fairly common, and small stumps that appear to be erect were found at the north base of the Hills. The occurrence of sandstone dikes at both ends of the area has already been mentioned. Foraminifera and fish scales indicating Kleinpell's Luisian stage (Kleinpell, 1938) were collected from fine-grained beds and partings in the western half of the Hills. Poor fossils from the eastern end suggest an uppermost Luisian or possible lowest Mohnian (Kleinpell, 1938) age, but, since no good lithologic change that might correspond to a Topanga-Puente contact was found, all the sediments were mapped as Topanga.

The volcanics in the Glendora South Hills are among the youngest known representatives of the Glendora volcanics. In terms of the local chronology, they are interbedded with late middle Miocene (Luisian and Mohnian(?)) sediments. Probably at least part of the subjacent sequence is present near by in The Texas Company's Covina 27–1 well a mile to the southwest (No. 8 on Pl. 1) which penetrated, in descending order, 480 feet of Quaternary deposits, 550 feet of Luisian sediments, 470 feet of interbedded sandstones and tuffs, and 3554 feet of predominantly andesitic breccias, tuff breccias, and massive units, the lower 1500 feet of which are somewhat albitized (Table 3).

The exposures a mile northeast of Azusa reveal about 2500 feet of bouldery sandstone and minor siltstone, if there is no duplication of beds in the area concealed by Quaternary deposits (section FG). In the southwestern part of the area 100 feet of basaltic lavas is interbedded with the sandstones. Luisian Foraminifera are reported from the eastern

TABLE 2.—CHEMICAL ANALYSES

Analysts: James Kerr (S-333$_2$, -339, -345, -347, -422, -438, -468, -496, -519, -597, -615)
Eileen H. Kane (Jones 1, S-669)
J. J. Engel (S-673)
M. K. Carron (Bayly 1)

	S-422	S-347	Jones 1, 1400 ft.	S-615	S-669	S-333$_2$	S-339	S-438	S-345	S-519	S-468	Bayly 1, 4000 ft.	S-597	S-673	S-496
SiO_2	47.23	48.20	48.75	51.06	57.14	59.53	59.90	62.85	63.19	64.93	66.16	68.75	72.16	73.10	75.50
Al_2O_3	16.65	16.89	15.30	15.36	14.64	18.30	17.89	16.78	18.18	17.20	16.65	14.76	13.73	13.74	13.45
Fe_2O_3	1.43	2.60	4.38	2.38	4.29	2.71	1.27	3.71	2.02	2.80	2.38	1.59	.32	.71	.63
FeO	7.66	6.86	1.79	5.85	.22	2.09	4.07	.22	1.47	.92	1.08	.83	.88	.35	.17
MgO	8.15	6.93	8.97	5.88	2.62	2.85	2.33	1.93	2.05	1.30	1.00	.95	.10	.14	.10
CaO	9.08	8.58	4.18	9.59	2.75	6.90	5.51	5.13	6.00	5.55	4.68	2.22	1.04	.69	.85
Na_2O	4.17	3.32	2.69	3.08	1.89	3.96	4.44	3.27	4.14	4.20	4.40	3.44	3.69	3.94	3.85
K_2O	.37	1.16	.36	.78	.46	.98	.90	.51	.71	.73	1.01	2.74	3.95	3.87	3.88
H_2O+	2.77	1.35	3.37	1.02	4.77	.47	1.56	2.16	.50	.73	.63	1.79	3.52	2.71	.62
H_2O-	.39	.80	8.90	.73	10.30	.88	.30	2.59	.90	.78	.66	2.06	.14	.36	.43
CO_2	.01	.09	.06	1.49	.01	.00	.01	.00	.00	.00	.00	.00	.01	.03	.01
TiO_2	1.59	2.33	.94	1.60	.74	.93	1.09	.61	.56	.53	.75	.22	.04	.03	.03
P_2O_5	.35	.49	.17	.72	.20	.28	.42	.25	.16	.28	.20	.00	.03	.02	.04
S	.03	.02	.04	.01	.02	.04	.01	.01	.03	.03	.03	.05	.02	.03	.13
MnO	.16	.14	.05	.12	.02	.06	.11	.01	.04	.03	.03	.18	.04	.02	.01
BaO	.02	.03	.00	.06	.00	.03	.01	.04	.03	.14	.03	.07	.07	.06	.14
	100.06	99.79	99.95	99.73	100.07	100.01	99.82	100.07	99.98	100.15	99.69	99.65	99.74	99.80	99.84
Less 0 for S:	.02	.01	.02		.01	.02			.02	.02	.01		.01	.01	.07
	100.04	99.78	99.93	99.73	100.06	99.99	99.82	100.07	99.96	100.13	99.68	99.65	99.73	99.79	99.77
Sp. G.*	2.826	2.847	....	2.788	....	2.600	2.630	2.438	2.595	2.557	2.576	....	2.332	....	2.389

C. I. P. W. Norms and Niggli Values

Sp. G.*	2.826	2.847	....	2.788	....	2.600	2.630	2.438	2.595	2.557	2.576	....	2.332	....	2.389

C. I. P. W. Norms and Niggli Values

Salic															
Q	....	....	9.0	5.6	34.8	15.9	14.0	28.9	20.7	14.6	25.4	32.3	32.1	33.7	36.4
or	2.2	6.7	2.2	4.5	2.8	5.6	5.6	2.8	4.4	22.8	6.2	16.7	23.3	22.8	22.8
ab	27.2	28.2	23.0	26.2	16.2	33.5	37.7	27.8	35.1	35.6	38.2	30.4	31.4	33.5	32.5
an	25.6	27.8	20.8	25.8	12.8	29.2	24.4	23.6	28.6	16.7	22.5	11.7	5.3	3.6	4.5
ne	4.3	....	....	....	....	....	....	....	....	....	....	....	....	....	....
C	....	....	2.7	....	6.2	....	0.5	2.2	....	....	....	1.8	1.3	1.6	1.3
Total	59.3	62.7	57.8	62.1	72.8	84.2	82.2	85.3	88.8	89.7	92.3	92.9	93.4	95.3	97.5
Femic															
ap	0.6	1.0	....	1.7	0.3	0.6	0.9	0.6	0.3	0.6	0.3	....	....	....	....
il	3.0	4.4	1.8	3.0	0.5	1.8	2.1	0.5	2.1	1.1	1.4	0.4	0.1	....	....
mt	2.1	3.7	3.0	3.5	....	3.9	1.9	....	3.0	1.4	1.4	2.3	0.5	0.9	0.2
di	14.2	9.1	....	6.0	....	....	....	....	....	....	....	....	....	....	....
ol	17.4	9.2	....	....	....	....	....	....	....	....	....	....	....	....	....
hy	....	7.1	....	18.5	....	7.2	10.8	4.8	5.1	3.2	....	....	1.6	0.4	0.2
cc	....	0.2	....	3.4	....	....	....	....	....	....	....	....	....	....	....
hm	....	....	2.4	....	4.3	....	....	3.7	....	1.9	1.4	....	....	....	0.5
en	....	....	22.4	....	6.5	....	....	....	....	....	2.5	2.4	....	....	....
wo	....	....	....	....	....	0.1	....	....	0.1	3.7	....	....	....	....	....
ru	....	....	....	....	0.5	....	....	0.4	....	....	....	....	....	....	....
Total	37.3	34.7	29.6	36.1	12.1	13.6	15.7	10.0	10.6	11.9	7.0	5.1	2.2	1.4	0.9
Niggli values															
al	22.4	24.1	26.1	24.8	40.9	35.0	35.8	40.1	38.8	39.6	40.9	44.4	48.7	47.5	50.2
fm	45.5	44.0	52.8	37.7	35.0	26.4	27.4	23.6	21.4	19.2	17.5	17.6	7.3	6.2	4.6
c	22.3	22.3	12.9	28.0	14.0	24.1	20.0	22.3	23.4	23.4	21.0	12.1	6.9	4.6	6.1
alk	9.8	9.6	8.2	9.5	10.1	14.5	16.8	14.0	16.4	17.8	20.6	25.9	37.1	41.7	39.1
si	108	117	148	139	272	194	204	252	230	253	277	352	440	430	477
k	0.06	0.18	0.08	0.14	0.14	0.14	0.12	0.09	0.09	0.11	0.13	0.34	0.41	0.35	0.40

TABLE 2.—*Continued*

	S-422	S-347	Jones 1, 1400 ft.	S-615	S-669	S-333$_2$	S-339	S-438	S-345	S-519	S-468	Bayly 1, 4000 ft.	S-597	S-673	S-496
C. I. P. W. Norms and Niggli Values—Continued															
Niggli values—Continued															
mg	0.62	0.57	0.73	0.64	0.53	0.53	0.43	0.49	0.52	0.40	0.36	0.41	0.15	0.20	0.25
ti	2.8	4.2	2.0	3.3	2.6	2.3	2.9	1.9	1.5	1.6	2.3	0.8	0.00	0.14	0.00
qz	−32	−23	15	3	132	36	38	96	64	81	95	148	192	163	221

* Apparent specific gravity.

S-422 Olivine basalt associated with Topanga sediments from foothills northeast of Azusa, California, 600 feet west and 1100 feet south of northeast corner of section 26, T. 1 N., R. 10 W., San Bernardino base and meridian.

S-347 Olivine basalt flow from upper east fork of Sycamore Canyon, north of San Dimas, California, 650 feet south and 1400 feet east of northwest corner of section 26, T. 1 N., R. 9 W., San Bernardino base and meridian.

Jones 1, from depth of 1400 feet, Los Angeles Brewing Company's Jones well No. 1, 856 feet south and 462 feet east of northwest corner of section 24, T. 2 S., R. 11 W., San Bernardino base and meridian.

S-615 Olivine basalt flow from 1000 feet southeast of summit 954, U. S. Geological Survey Covina 6-minute quadrangle, northwest of Puddingstone Dam, Los Angeles County, California.

S-669 Andesite tuff in road cut on Gladstone Avenue at west end of Way Hill, San Dimas, California, 5500 feet west and 5100 feet north of southeast corner of U. S. Geological Survey Glendora 6-minute quadrangle.

S-333$_2$ Calcic andesite flow from upper Wildwood Canyon, 2600 feet south and 1425 feet west of northeast corner of section 27, T. 1 N., R. 9 W., San Bernardino base and meridian.

S-339 Fine-grained andesite flow from east side of lower Mull Canyon, 2950 feet south and 950 feet west of northeast corner of section 28, T. 1 N., R. 9 W., San Bernardino base and meridian.

S-438 Matrix of Johnstone Peak tuff breccia from north of San Dimas Foothill Park, 625 feet west and 150 feet south of center of section 26, T. 1 N., R. 9 W., San Bernardino base and meridian.

S-345 Hypersthene andesite block in Johnstone Peak tuff breccia, from north of San Dimas Foothill Park, 600 feet west and 100 feet south of center of section 26, T. 1 N., R. 9 W., San Bernardino base and meridian.

S-519 Block in hornblende andesite breccia of Harwood Hills, San Dimas, California, 750 feet N. 77° E. from triangulation station 1253, U. S. Geological Survey Claremont 6-minute quadrangle.

S-468 Block in hypersthene andesite breccia, south of Mountain Meadows Country Club, Pomona, California. About 6200 feet east and 9100 feet south of northwest corner of U. S. Geological Survey Claremont 6-minute quadrangle.

Bayly 1, from depth of 4000 feet, Continental Oil Company well Bayly No. 1, about 3050 feet north and 2570 feet east of southwest corner of section 29 (projected), T. 1 S., R. 9 W., San Bernardino base and meridian. Near Buzzard Peak, about 3 miles southeast of Covina, California.

S-597 Pocket of perlite in felsite breccia, 400 feet south-southeast of summit 1009, 0.6 mile west of Puddingstone Dam, Los Angeles County, California. About 6700 feet west and 4500 feet south of northeast corner of U. S. Geological Survey Covina 6-minute quadrangle.

S-673 Glassy block in Elephant Hill mass, a few hundred feet southeast of triangulation station 1160, about 1500 feet east and 1770 south of northwest corner of U. S. Geological Survey Clarement 6-minute quadrangle. About 2½ miles west of center of Pomona, California.

S-496 Felsite from streambed at northwest base of Hill 1063, 1 mile west of La Verne, California. About 750 feet south and 4100 feet east of northwest corner of U. S. Geological Survey Claremont 6-minute quadrangle.

of U. S. Geological Survey Claremont 6-minute quadrangle. About 2½ miles west of center of Pomona, California.

S-496 Felsite from streambed at northwest base of Hill 1063, 1 mile west of La Verne, California. About 750 feet south and 4100 feet east of northwest corner of U. S. Geological Survey Claremont 6-minute quadrangle.

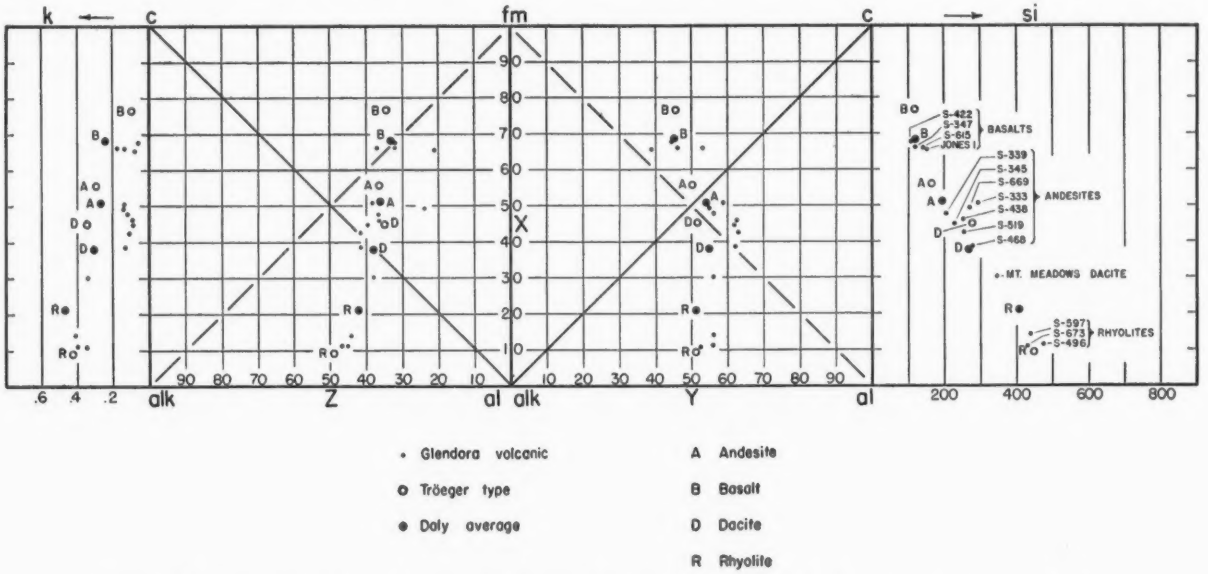


FIGURE 3.—NIGGLI-BECKE-BACON DIAGRAM COMPARING CHEMICAL COMPOSITIONS OF GLENDORA VOLCANICS WITH TRÖGER'S TYPES AND DALY'S AVERAGES

part of the area, and the highest beds exposed yielded fish scales that are probably either uppermost Luisian or lowermost Mohnian like those in the east end of the Glendora South Hills; thus these two areas probably expose about the same part of the section. The five small exposures labelled Topanga(?) in the foothills west of San Gabriel Canyon have yielded no diagnostic fossils, but their lithology resembles the Topanga beds north of Azusa.

Five miles west of Pomona an asymmetrical anticline brings to view a thick conglomerate along the crest of the San Jose Hills. In the Ohio Legrand 1 well (No. 47 on Pl. 1; Pl. 6, section RST) this interfingers with finer sediments containing Luisian Foraminifera and may accordingly be referred to as the Buzzard Peak member of the Topanga formation (Woodford *et al.*, 1946, p. 515, 518). At least half the volume of the rock is arkosic sandy matrix; the pebbles and boulders reach 3 feet in diameter but sizes near 1 inch predominate. The latter are not very well rounded except for the mylonites, and include only about 10 per cent volcanics. Maximum thickness based on surface and well data is probably about 3000 feet.

A layer of much-altered volcanics, less than 15 feet thick, occurs along much of the Topanga-Puente contact around Buzzard Peak. Some of the rock is fine-textured andesite, some is near the andesite-basalt line, and much is altered beyond recognition. It is interpreted as another late submarine accumulation of Topanga age, possibly separated from the Puente deposits by part of the unconformity which, 1 mile east, places Puente directly on the main mass of the volcanics (Pl. 6, section RST).

Puente Formation

As now used the Puente formation, first defined by Eldridge in 1907, includes all the upper Miocene rocks on the east side of the Los Angeles Basin. It is thus approximately equivalent to the Modelo formation of the Santa Clara Valley and is one of the southern California representatives of the Monterey shale (Bramlette, 1946).

Within the area covered by Plate 1 it is typically composed of siltstone and shale, some of which is of the siliceous platy type so characteristic of the Monterey, but it also includes thick local lenses and tongues of sandstone and conglomerate, especially in the upper part and close to its northeastward limits of outcrop. Calcareous concretions, impure limestone lenses, and fragments of silicified wood are locally abundant in the lower part.

As already pointed out the lower 2000 feet or so of the Puente formation laps out southward against the volcanics between the Puddingstone area and Elephant Hill. In the foothills, especially between Shuler and Shay canyons, hornblende andesite breccia and other volcanics have been reworked to form a local overturned basal conglomerate 5 to 50 feet thick composed of smooth rounded pebbles and cobbles up to 15 inches in diameter. Foraminifera from the overlying thin-bedded sandy shales in this area and southwest of Puddingstone reservoir belong to the lower part of Kleinpell's Mohnian stage (lower upper Miocene by the locally accepted marine invertebrate standards of correlation); the Puente sediments resting on the volcanics of the Spadra area are probably a little younger.

No Puente sediments have been identified in outcrops along the margin of the Basin west of Dalton Canyon, but some water wells have encountered them (Table 3).

The Puente formation reaches a maximum thickness of about 9000 feet in the Puente Hills; the thickest part within the area of Plate 1 probably occurs southwest of Spadra where it is inferred to reach 5000 feet; in the foothills north of San Dimas the exposed thickness is about 2000 feet.

Duarte Conglomerate

The name Duarte conglomerate has been proposed (Shelton, 1946) for the coarse, unsorted sandy conglomerate exposed in isolated outcrops showing through Quaternary deposits in the foothills between Sawpit and Maddock canyons north of Duarte (Pl. 1, inset). Here it is characterized by steep dips, poor consolidation resulting in rarity of good natural outcrops, and a light-gray color in contrast with the yellow, buff and reddish brown conglomerates of the Topanga, Puente, and San

Dimas formations. The boulders and pebbles are predominantly of basement rock types; gneisses and biotite plutonites are common, while aplite, pegmatite, and volcanic rocks are rare. A few boulders of sandstone resembling the Topanga were found. The largest boulder seen in place was 33 inches across; the largest as float, 5 feet. All are quite well rounded and generally show little or no sorting or preferred orientation. The sandy matrix is estimated to constitute 20 to 40 per cent of the volume of the rock.

The best exposures are small road cuts west of the mouth of Bradbury Canyon in the southwest quarter of section 19, T. 1 N., R. 10 W. Here many of the boulders are so decomposed they have been cut through by the bulldozer, and occasional beds of sand and fine gravel show graded bedding with dips of 50°–80° N. and repeated evidence that these are overturned.

The maximum exposed thickness of the Duarte conglomerate is at least 1500 feet, due north of Duarte (Pl. 6, sections CE and CD). No fossils have yet been found in it. Since it is relatively unconsolidated, seems to overlie the Topanga(?), and is very different lithologically from anything known in the Miocene of this region and yet has overturned dips and lies unconformably beneath the Pleistocene San Dimas formation, it is tentatively assigned to the Pliocene. It is probably a fanglomerate and may be the subaerial equivalent of part of the thick marine Pliocene section exposed 11 miles southwest. The possibility that it bears the same relationship to late Miocene deposits of the Basin cannot be eliminated, though it is considered remote.

San Dimas Formation and Terrace Deposits

Since it was first pointed out by Mendenhall (1908), local geologists have recognized two ages of alluvium in this region, the gray fresh and uncemented "later alluvium" of the present cycle and an "earlier alluvium" of more or less decomposed gravel in a clayey matrix whose iron-oxide stain and cement give it characteristic reddish and brown colors that are also reflected in its soils. This older alluvium Eckis (1928) included in his San Dimas formation.

The San Dimas formation is extensively exposed as remnants of Pleistocene fan deposits along the base of the San Gabriel Mountains where in many places it is clear that it was deposited on an irregular surface. Samples from scores of water wells show that the gray alluvium is a relatively thin veneer beneath which the more decomposed older alluvium of the San Dimas formation is much more extensive than it is in outcrop. Both in outcrop and in wells it is characteristically composed of a few small angular pebbles and cobbles of durable quartz-rich rock types (aplite etc.) and soft and rotten cobbles of gneiss and plutonic rocks in a red matrix notable for its high (50 per cent or more) clay and low sand content (Eckis, 1934). In some wells clay-rich zones alternate with less-decomposed ones. Eckis concluded that most of the clay was residual, having formed above the water table from the decomposition of basement complex boulders, and that each red or brown zone records a zone of oxidation associated with a particular stage and position of the water table during the growth of the fans.

Well data show that the continental San Dimas formation lies on basement complex and marine Tertiary deposits at depths in some places exceeding 1000 feet and not uncommonly below sea level. Its age is based on several occurrences of fossil *Elephas imperator*(?), such as the one found 60 feet below the surface near San Dimas (Eckis, 1928), and on the fact that it clearly formed in an earlier cycle. Although it is nearly parallel to the present surface over large areas, it has dips of 35° to 83° close to the range-front fault system between Monrovia and Duarte.

As here mapped, the San Dimas formation includes some terrace remnants in the foothills that probably are not strictly comtemporaneous with the red-brown alluvium of the lowlands. Some of the more obscure and gradational boundaries, such as that passing just west of La Verne, are taken from unpublished maps by Eckis who based them in part on soil samples.

Structure

Two broad structural patterns impinge on each other in this area. The characteristic

structures within the Los Angeles Basin sediments are gentle to moderate folds, many of them faulted at depth. In the northeastern part of the Basin (Puente and San Jose Hills) most of these trend northeastward or northwestward, but close to the San Gabriel Mountains this pattern is overshadowed by the dominant east-west trends of the piedmont and foothills belt, as though this margin of the Basin had been compressed and broken by pressure from the north.

Thus, in general, the intensity of folding decreases southward away from the base of the mountains (Pl. 6). The steep and overturned beds occupy only a narrow belt in the foothills; 1–2 miles south, as in the Glendora South Hills, the average dip is less than 30°. Here, as in the foothills belt, there is still a strong parallelism between the strikes in the sediments and the trend of the mountain front. Farther south, in the San Jose Hills, the larger folds trend northeasterly, and there is enough difference between the deformation in the sediments and volcanics to suggest that the rock type may have exerted some influence on the result.

There are undoubtedly many more faults in the area than are shown on Plate 1; in general, only the more conspicuous ones involving contacts were mapped. The few shown entirely within the basement complex are based largely on topographic and groundwater evidence. In the foothills the Tertiary rocks are usually separated from the basement by one of the many faults that together make up the frontal fault system of the San Gabriel Range. Topographic expression and rare exposures suggest that all are steep and probably most dip toward the mountains. One of the best exposures of an actual plane of movement (1.6 miles northeast of Azusa) reveals a red clay gouge zone 12 inches thick dipping 75° N., north of which the basement complex is notably sheared and shattered for more than 100 feet. Just east of the mouth of San Gabriel Canyon Topanga sandstone and volcanics are separated from basement complex by a fault that strikes N. 80°E. and dips 64° N.; the exposure is a large excavation. Such steep reverse faults are probably the habit of this belt. As is sometimes the case in the Basin and Range province (as for example at the west base of the Panamint Range), not all the frontal faults are upthrown on the mountain side; each contrary individual produces a graben and horst. Small examples may be seen in the western part of the mapped area.

The belt labelled "Zone of faulting and sliding" (Pl. 1) is one in which the few discontinuous contacts that can be followed do not make structural sense, on the scale of present mapping, except as part of such a zone. Units that are fairly continuous near by become pinched off, minor faulting is abundant, and some of the members of the volcanics are difficult to distinguish. The zone lies along and beyond the foot of a scarp (Pl. 5) which is probably pre-Pleistocene, as the gently rolling low-relief topography topping the lower block is 100 to 200 feet above the San Dimas formation and is deeply dissected by many canyons. The details of structure shown beneath this zone on section MN are highly conjectural. Evidence from water wells supports the possibility that this zone may help to produce an extremely rapid thickening of the alluvium just below the mouth of Big and Little Dalton canyons.

Several faults shown under the alluvium (Pl. 1) are based chiefly on ground-water levels in wells. The demonstrable Quaternary faults in this region commonly impede the downslope movement of ground-water with resulting higher levels north of the fault and somewhat independent fluctuation of the water table on the two sides. By analogy, if it is geologically reasonable, queried faults are shown on Plate 1 where the difference in average water levels, with or without other evidence, is the chief clue. The San Jose fault between Pomona and Claremont might be suspected from a map of well locations alone, without knowing what they encountered, for along this segment an unpublished well-location map of the State Division of Water Resources shows more than 60 wells within 2000 feet of the fault on its northwest side and only 8 (at least 1 of which was dry) within a mile of it on the southeast side. In this case evidence from a few deep wells indicates that the southeast side moved relatively down; however, the water table would probably be similarly affected here by any reasonable combination of horizontal and vertical movements.

Water-well data collected by Eckis (1934)

show that the surface of the basement complex close to the San Gabriel Mountains is one of irregular ridges and canyons that drained southeastward (probably to the Santa Ana River) in the La Verne-Claremont area and southwestward in the Glendora-San Dimas area. Extension of gradients for the bedrock surface by means of wells drilled for oil indicate an average slope southwestward beneath the axis of the San Jose Hills of 10°–11°, which is a minimum because the direction of maximum slope is not accurately known. The slope between the Covia 27–1 well, which may not have reached basement, and the nearest outcrops is at least 15°–16°, some of which is accomplished by step faulting.

Geomorphology

Three stages in the late Pliocene(?) and Quaternary history of the area are suggested by the existing landforms. The earliest is recorded in areas of subdued broad-valley topography such as that between Dalton and San Dimas canyons south of the "zone of faulting and sliding" (Pl. 1), or by the triangular area of basement complex exposed between Marshall Creek and Liveoak Canyon north of La Verne (Pl. 5). In both these areas many elevations are within 100 feet of 1500 feet, and the gently rolling topography with probable erstwhile west-southwest flowing drainage contrasts with the rugged mountain front to the north and the escarpment by which these areas are separated from the alluvial plain to the south. Such canyons as Gordon, Mull, Morgan, Wildwood, Shuler, Shay, Sycamore, and Liveoak have cut sharply into this surface, slicing through the old drainage lines and developing precipitous walls. The small streams that drain the old surface exhibit marked convexities in their long profiles where they leave it to plunge into these later canyons. A small patch of similar topography near 1100 feet in the Mountain Meadows-Ganesha Park area may belong to this same stage. These surface remnants are believed to be older than the San Dimas formation because in different areas the latter either was apparently deposited on them (southeast of Marshall Creek) or is confined to lower elevations. The Liveoak Canyon remnant may now be an exhumed surface.

The second stage is recorded in the building of the piedmont alluvial slope whose remnants we now call the San Dimas formation of Quaternary (probably upper Pleistocene) age. Within much of the area covered by Plate 1 this constructional surface was either graded to a higher base level than the modern fans or the remnants have been uplifted, or both; in any case warping has taken place in many areas.

The most recent stage is the growth of the present-day fans following or concurrent with the removal of some of the San Dimas deposits.

Age of the Glendora Volcanics

Some of the best evidence for the age of the exposed volcanics comes from the western two-thirds of the Glendora South Hills where four collections of Foraminifera and eight of fish scales all indicate Luisian (upper middle Miocene) age. One of the former is a lapilli tuff interbedded with the sandstones, collected by Bradley Sauters and containing very well preserved *Siphogenerina nuciformis*, *S. reedi*, and *Valvulineria californica* var. *obesa*—all good guides to the upper middle Miocene. For reasons already cited the interbedded volcanics here are believed to be among the youngest representatives of the series. In Way Hill a sliver of shale embedded in the volcanics near the summit yielded scales from 11 genera of fish interpreted by Dr. Lore R. David to be not younger than upper Luisian, while mollusks and echinoids from a small exposure of coarse sandstone on the north flank are considered middle Miocene by William H. Corey (personal communication). This sandstone is probably interbedded with the volcanics.

Near the Rainbow Club, northeast of Azusa, Luisian Foraminifera are reported from the sandy shales in which occur 100 feet of basaltic lava; this occurrence is considered comparable to that in Glendora South Hills. In the Glendora foothill area, at the east end of Sierra Madre Avenue, a concretion in the sandstone and shale in the creek bed yielded a *Valvulineria californica* fauna representing Kleinpell's

Luisian stage. The relation of these sediments to the volcanics is not clear; they are probably interbedded with them.

Within the area covered by Plate 1, five wells (Nos. 1, 8, 46, 47, 48) went from Luisian or Relizian(?) (middle Miocene) sediments into the volcanics (Table 3).

In all the other areas of Plate 1 where the contact is not faulted the volcanics are overlain by Puente sandstones and shales containing Foraminifera of lower Mohnian (lower upper Miocene) age, usually the lowest Mohnian *Bolivina modeloensis* zone. Numerous fish-scale collections give the same results. At most places there is evidence that the Puente-volcanics contact is an unconformity.

The evidence at these many localities is all consistent with a middle Miocene age for all the Glendora volcanics in the area mapped.

Interpretive Geology

Extent of Probable Correlatives of the Glendora Volcanics

Subsurface.—Figure 1 shows the locations of 39 wells drilled for oil, selected to indicate the distribution and age of the volcanics. The data on these wells are summarized in Table 1. (The geological information obtained from the wells shown on Plate 1 is summarized in Table 3.) This evidence strongly suggests that the Glendora or closely equivalent volcanics extend beneath most of at least the eastern half of the Los Angeles Basin. Westward from the Santa Ana Mountains outcrops they are found beneath a thickening cover of Miocene and Pliocene marine sediments. Along a narrow zone extending approximately from the northwest tip of the San Joaquin Hills to the Repetto Hills there is a marked steepening of westward dips. The volcanics probably simply pass out of reach west of this zone, for on the western side of the Basin similar rocks of similar age appear in some of the deeper wells in the Dominguez and Inglewood oil fields. (*See also* Schoellhamer and Woodford, 1951.)

In places in the shallower eastern part of the Basin well data demonstrate a post-Miocene unconformity along which the upper Miocene Puente formation is missing and Pliocene beds rest directly on the volcanics or lower horizons (Table 1). Also present in some of these wells are pre-volcanic sediments not seen in outcrops farther northeast. Likewise, many wells near the coast encounter incomplete sections and basement highs, again recording marginal unconformities resulting from the fluctuating size and shape of the Basin. The absence of the volcanics in some of these areas is probably the result of erosion during the interval represented by the unconformity.

The somewhat meager evidence for age of the volcanics in the central and western part of the Basin indicates that most of them embrace about the same interval of time as the Glendora volcanics. Of the 39 wells shown in Figure 1, three widely separated ones (Nos. 4, 5, and 9) reported middle Miocene Foraminifera in thin stringers of sediment within the volcanics. One (No. 35) may have a few feet of upper Miocene volcanics, but aside from this and some fine-grained ash falls, no undisputed evidence of an age other than middle Miocene is known to the writer. The volcanics in wells are usually overlain by Luisian or Mohnian sediments.

Petrographically the volcanics in well cores are predominantly fragmental, commonly tuffaceous, and usually andesitic or basaltic. Nearly all are somewhat altered; albitization is common in the lower part of the Covina 27-1 well (No. 8 on Pl. 1), sulphides are present in some, and the groundmass has usually been altered almost beyond recognition. There is no reason to believe they differ significantly, as a group, from the exposed Glendora volcanics. Most of the differences could be the results of subaqueous accumulation.

Surface.—Figure 1 also shows the known outcrop areas, somewhat generalized, of Miocene volcanic rocks, all of which are believed to be middle Miocene. Locally these have been named the Conejo volcanics in the Conejo Hills (Taliaferro *et al.*, 1924, p. 800) and the El Modeno volcanics at the west base of the northern Santa Ana Mountains (Schoellhamer *et al.*, 1954). Distinctions between intrusive and extrusive material are uncertain in some areas, especially the western Santa Monica Mountains and Conejo Hills. In some places sediments (not shown on Figure 1) are locally interbedded with the extrusives.

The thickest and most widespread of these probable correlatives of the Glendora volcanics occur in the western Santa Monica Mountains. For example, in the Seminole quadrangle (near Agoura) on the north flank Elam (1948) reports a northward-dipping sequence of over 14,000 feet of flows and tuffs that are predominantly basaltic and andesitic but include some dacitic tuff. They are at least locally overlain by fossiliferous Topanga sediments containing Relizian (Kleinpell, 1938) and possibly Luisian microfossils. Field relations and small lenses of interbedded sediment led Elam to interpret nearly all as submarine accumulations. In the Conejo Hills Taliaferro (1924) estimates a thickness of the order of 10,000 feet of flows, "agglomerates," and tuffs of probable middle Miocene age and largely submarine accumulation. In ascending order these consist chiefly of hornblende and augite andesites and hypersthene and olivine basalts belonging to the upper part of the Topanga formation (Thomas L. Bailey, personal communication). Forty miles farther west, in Santa Cruz Island, Bremner (1932) reports more than 5000 feet of rhyolitic, basaltic and andesitic flows and breccias between the fossiliferous Temblor (middle Miocene) formation and siliceous Monterey shales (middle and/or upper Miocene).

In short, the upper middle Miocene age of the later members of the volcanic series is indicated for many parts of the Los Angeles Basin. (Minor fine-grained tuffs interbedded with upper Miocene marine sediments are not considered part of the Glendora volcanics and its correlatives.) Concerning the earliest eruptions less is known. Around the margins of the Basin where they are exposed they rest on basement complex, and out in the Basin many wells stop drilling when the volcanics are encountered. In the northern Santa Ana Mountains, clearly on the edge of the Basin, no extrusive volcanics are known from the apparently nearly complete sequence of sediments beneath the Topanga formation and its associated volcanics down to the Upper Cretaceous.

Probable volume.—Clearly the exposed Glendora volcanics are but a small sample of widespread middle Miocene volcanism in Southern California. The outcrops shown in Figure 1 total over 125 square miles of extrusives and over 20 of intrusives. If essential continuity of the volcanics is assumed around the group of wells east of Oxnard (except No. 25, which bottomed in Sespe beds and may indicate the limit of the volcanics in that direction) and again those grouped on the east and west sides of the Los Angeles Basin, about 500 square miles is added to the area of exposures. If the deep central part of the Basin, in which the likely horizons are out of reach of current drilling, and much of the San Fernando valley are also occupied by volcanics, the total extent is of the order of 1000 square miles.

Maximum thicknesses (known in only a few areas) are: 8000–14,000 feet in the Santa Monica Mountains and Conejo Hills, at least 2000 feet in the exposed Glendora volcanics, and approximately 2990–3720 feet in three wells in and near the San Jose Hills (Nos. 8, 46 and 47, Pl. 1 and Table 3). Five wells shown on the east side of the Basin (Table 1, nos. 3, 15, 17, 32, 34) and one on the west side (No. 23) are known or believed to have gone through the volcanics into older rocks. The smallest penetration of the six was about 980 feet; the average 2370 feet.

This is obviously slim evidence on which to speculate regarding volume, but perhaps the order of magnitude is indicated by assuming an average thickness of 1000 feet over an area of 700 square miles which would give approximately 140 cubic miles. Because in several places the volcanics have been deeply eroded this figure may be a minimum, yet it is substantially larger than the San Franciscan volcanic field of northern Arizona (Robinson, 1913).

The Source Problem

The nature of the Glendora volcanics suggests at least one fairly large volcano as a source for the rocks interpreted as deposits of peléan eruptions, but little has been observed that might indicate in what direction such a vent lay from the present exposures. If such a vent was the only source, its ejectamenta must have covered at least 60 square miles and subsequent deformation and erosion should expose litho-

TABLE 3.—WELLS AND COREHOLES AFFORDING RELIABLE INFORMATION ON SUBALLUVIAL GEOLOGY

Note: All wells shown on Plate 1 are listed in this table. The water well numbers are the "location numbers" shown on maps in the Los Angeles office of the California State Division of Water Resources. Rollin Eckis, James C. Kimble, and Clifton W. Johnson, geologists employed by the Division of Water Resources, observed these wells (except No. 4397A) during drilling and collected rock samples many of which are preserved at Pomona College, Claremont, Calif. Many water wells not listed here are reported to have reached shale beneath Pleistocene deposits.

Map No.	Well	Drilled for	Approximate location (*see* geologic map)	Year drilled	Elev. (feet)	Total depth (feet)	Remarks (Figures are depths in feet)
1	Covina corehole No. 4		2.2 mi S. of Glendora	1943	695	1086	0–430 Quaternary gravels 430 –1007 dark-brown and gray shale with Foraminifera and fish scales; streaks of sand and local chert; dips 6°–14°; 11 cores. Reported Luisian* below 704 1010(?) –1086 volcanics; 2 cores
2	Covina corehole No. 5		2.4 mi S. of Glendora	1943	685	1090	0–430 Quaternary gravels 430 –1090 shale, siltstone, and sandstone; Foraminifera, fish scales, and mollusks; phosphatic nodules; dips chiefly 7°–16°; 6 cores. Reported Luisian* below 470
3	Covina corehole No. 7		2.1 mi W. of San Dimas	1943	755	955	450 – 760 sandy dark shale, with occasional Foraminifera and fish scales; dips 10°–12° in upper part; broken and slickensided at 750; 4 cores
4	Covina corehole No. 8		2.0 mi W. of San Dimas	1943	765	995	0–300 Quaternary gravels 315 – 510+ brownish-gray sandy siltstone; fish scales and abundant Foraminifera; phosphatic nodules; dip 23°; 2 cores. Reported bottomed in Luisian*-lower Mohnian* transition
5	Covina corehole No. 9		2.1 mi W. of San Dimas	1943	775	919	310 – 400± conglomerate; 1 core 440 – 720 sandy shale with Foraminifera and fish scales; phosphatc nodules; dips 18°–20°; 2 cores
6	Covina corehole No. 10		2.4 mi S. of Glendora	1943	640	1075	540 –1060 sandy siltstone, shale, and sandstone; Foraminifera and fish scales; phosphatic nodules; dips 13°–21°; 6 cores. Reported Luisian* below 540

										nodules; dips 13°–21°; 6 cores. Reported Luisian* below 540

7	Covina corehole No. 13		2.4 mi S. of Glendora	1943	?646	1060		585 – 955		silty sandstone; dip 17°; 4 cores
								1050 –1060		vesicular calcic andesite, 1 core
8	Covina 27–1	The Texas Co.	2.3 mi S. of Glendora	1944–45	665	5054	Condensed log, based primarily on cores (641 feet cored, 407 feet recovered); contacts marked (E) estimated from electric log:			
							0 – 480			Quaternary alluvium and San Dimas formation and terrace deposits. No cores
							480 –1030(E)			Luisian* sediments (Topanga (?) formation). No cores
							1030(E)–1500			sandstone, tuff, and tuffaceous sandstone; megafossils. 13 cores
							1500 –2026			andesite (?) and basaltic tuff and sandy tuff. 8 cores
							2026 –2060			calcic andesite. 4 cores
							2060 –2450(E)			predominately tuff breccia. 4 cores. Dip 15° at 2292 (tuff)
							2450(E)–2975			massive and brecciated basalt. 11 cores
							2975 –3300(E)			massive and brecciated andesite. 7 cores
							3300(E)–3560(E)			andesitic breccia and albitized basaltic breccia. 5 cores
							3560(E)–4225(E)			albitized, massive (?) andesite. 8 cores
							4225(E)–4690(E)			albitized massive basalt. 10 cores
							4690(E)–5054			mica-albite rhyolite porphyry. 3 cores
9	Water well No. 4325D	Charles Richardson	1.0 mi NW. of Glendora	1931	892	365	Entered sandstone and conglomerate at 360 feet			
10	Water well No. 4359	Oak Water Co.	2.1 mi W. of San Dimas	1930	765	324	0– 240±	Quaternary gravels	240±– 324	shale and sandy shale
11	Water well No. 4368	Columbia Water Co.	1.6 mi W. of San Dimas	1930	823	316	0– 208	Quaternary gravels	208 – 290	shale and sandstone
									290 – 316	volcanic breccia
12	Water well No. 4376E	J. H. Moore	1.7 mi NW. of San Dimas	1929	895	362	0– 250±	Quaternary gravels	250±– 362	dark shale of the Puente (?) formation
13	Water well No. 4378B	C. R. Luton	1.1 mi W. of San Dimas	1931	858	96	0– 30±	Quaternary gravels	30±– 96	sandstone and siltstone
14	Water well No. 4379A	C. R. Luton	0.7 mi W. of San Dimas	1931	864	382	0– 350	Quaternary gravels	350 – 382	sandstone and sandy shale
14a	Water well No. 4397A	San Dimas Water Co.	0.5 mi N. of San Dimas	1919	1008	517			415 – 517	shale (J. C. Kimble, who saw samples)
15	Water well No. 4398A	San Dimas Water Co.	0.5 mi NE. of San Dimas	1930	1018	240	0– 185±	Quaternary gravels	185±– 188	weathered volcanic rock
									188 – 220	gneiss
									220 – 226±	volcanic breccia
									226±– 238	sandy shale. 30° (?) dip

TABLE 3.—*Continued*

Map No.	Well	Drilled for	Approximate location (*see* geologic map)	Year drilled	Elev. (feet)	Total depth (feet)	Remarks (Figures are depths in feet)			
16	Water well No. 4407T	Gillette Bros.	1.7 mi NE. of San Dimas	1931	1092	298	0– 205+	Quaternary gravels	240 290	conglomerate (?) sandy shale with Luisian* Foraminifera (Bramlette)
17	Water well No. 4416H	D. R. Malone	2.5 mi NE. of San Dimas	1930	1275	762	0– 186	Quaternary gravels	300(?)– 418 418 – 704– 704–– 750	andesite (?) breccia dacite (?) breccia sandy shale and arkose
18	Water well No. 4416I	San Dimas Water Co.	2.2 mi NE. of San Dimas	1931	1155	206	0– 186	Quaternary gravels	186 – 206	gray shale like that in well No. 4416H
19	Water well No. 4425B	Budd Gray	3.2 mi NE. of San Dimas	1931	1375	502	0– 295±	Quaternary gravels	295±– 502	basement (gneiss or schist)
20	Water well No. 4426B	D. R. Malone	2.8 mi NE. of San Dimas	1931	1215	460			270 – 460	dark-gray or brown fissile, somewhat sandy shale, with good lower Mohnian* Foraminifera (Bramlette)
21	Water well No. 4429D	W. A. Hays	0.6 mi N. of La Verne	1932	1078	347	0– 204	Quaternary gravels	204 – 245 245 – 347	volcanic rocks dacite porphyry
22	Water well No. 4439E	Bonita Water Co.	In La Verne	1932	1060	757	0– 700	Quaternary gravels(?)	700 – 757	biotite dacite porphyry
23	Water well No. 4439H	Evergreen Ranch Co.	N. edge of La Verne	1931	1072	416	0– 398	Quaternary gravels	398 – 416	disintegrated dacite porphyry
24	Water well No. 4448B	R. K. Pitzer	1.5 mi NE. of La Verne	1931	1255	471	0– 400	Quaternary gravels	400 – 471	basement (gneiss and pegmatite)
25	Water well No. 4458E	La Verne Water Co.	1.8 mi NE. of La Verne	1931	1285	530	0– 520	Quaternary gravels	520 – 530	basement (gneiss)
26	Water well No. 4479E	North Palomares Water Co.	0.8 mi NW. of Claremont	1932	1223	790	0– 735	Quaternary gravels	735 – 790	gneiss
27	Water well No. 3194H	Graffen Bros & Hicks	Spadra	1931		186	0– 180	Quaternary gravels	180 – 186	shale

27	Water well No. 3194H	Graffen Bros & Hicks	Spadra	1931		186	0– 180	Quaternary gravels	180 – 186	shale
28	Water well No. 3195L	Goodell	0.3 mi E. of Spadra	1931	712	198	0– 150	Quaternary gravels	150 – 190	volcanic breccia
									190 – 198	biotite dacite porphyry
29	Water well No. 3195N	E. C. Basten	0.4 mi E. of Spadra	1931	721	97	0– 87	Quaternary gravels	87 – 97	volcanic breccia
30	Water well No. 3210	Baseline Water Co.	1.1 mi E. of Puddingstone reservoir	1930	981	440	0– 352±	Quaternary gravels	352±– 440	felsitic rhyolite
31	Water well No. 3210B	Baseline Water Co.	1.3 mi E. of Puddingstone reservoir	1930	993	405	0– 403	Quaternary gravels	403 – 405	felsitic rhyolite
32	Water well No. 3213C	Phillips Estate	2 mi W. of Pomona	1931	796	264	0– 264	Quaternary gravels	(clayey, with gypsum and volcanic pebbles in lower part)	
33	Water well No. 3220B	Evergreen Ranch Co.	0.5 mi S. of La Verne	1930	1017	750	0– 745	Quaternary gravels	745 – 750	gneiss
34	Water well No. 3223	Orange Grove Tract Water Co.	1.3 mi W. of Pomona	1931	833	810	0– 356	Quaternary gravels	356 – 810	gneiss
35	Water well No. 3223B	Calif. Fruit Wrapping Mills, Inc.	1.4 mi W. of Pomona	1930	823	350	0– 332	Quaternary gravels	332 – 350	decomposed Mt. Meadows dacite(?)
36	Water well No. 3224F	Phillips Estate	1.5 mi W. of Pomona	1931	812	248	0– 248	Quaternary gravels		
37	Water well No. 3230B	San Dimas Water Co.	0.3 mi S. of La Verne	1930	1036	900	0– 882	Quaternary gravels	(chiefly volcanic around 830-850)	
									882 – 900	Mt. Meadows dacite
38	Water well No. 3230G	Valentine Peyton Corp.	0.1 mi S. of La Verne	1932	1048	870	0– 846	Quaternary gravels	846 – 870	decomposed gneiss
39	Water well No. 3233	Currier Tract Water Co.	0.9 mi W. of Pomona	1931	850	382	0– 368	Quaternary gravels	368 – 382	gneiss
40	Water well No. 3251A	Live Oak Water Co.	2 mi N. of Pomona	1930	1047	994	0– 955±	Quaternary gravels	955±– 994	gneiss

TABLE 3.—*Continued*

Map No.	Well	Drilled for	Approximate location (*see* geologic map)	Year drilled	Elev. (feet)	Total depth (feet)	Remarks (Figures are depths in feet)
41	Water well No. 3251B	City of Pomona	1.6 mi N. of Pomona	1930	1002	1104	0–1104 Quaternary gravels
42	Water well No. 3252F	Irrigation Co. of Pomona	1.2 mi N. of Pomona	1931	958	690	0– 690 Quaternary gravels
43	Water well No. 3260A	Avenue Line Mutual Water Co.	1.2 mi W. of Claremont	1930	1133	780	0– 668 Quaternary gravels 668 – 780 Mt. Meadows dacite
44	Water well No. 3264A	Irrigation Co. of Pomona	E. edge of Pomona	1931	887	1107	0–1100 Quaternary gravels 1100 –1107 gneiss
45	Water well No. 3271C	Adams & Garner	0.9 mi SW. of Claremont	1931	1083	740	0– 672 Quaternary gravels 672 – 740 basement complex (diorite?)
46	Garnier 1	The Texas Co.	3.3 mi S. of Covina	1944–45	1180	7148	0 –3060 Upper and middle (?) Miocene siltstone, shale and sandstone. Lower Mohnian* in upper part. 1600 –2251 reported as Luisian* 2450 –2726 reported as Relizian* 3060 –6780 Glendora volcanics (base estimated from electric log) 3060 amygdaloidal augite andesite, pyritized 3141 feldspathic basalt 3158 ?albitized hornblende andesite(?) with laumontite 3168 altered hornblende andesite(?) with laumontite 3817 –3822 andesitic tuff 3876 albite porphyry (spilite?) 3892 albite porphyry (spilite?) 3930 tuff breccia 4360 breccia 5160 fine-grained porphyritic augite andesite 5444 breccia 6130 tuff or tuff breccia 6515 altered fine-grained andesite flow (like sample from 5160) 6780 –7148 Basement complex 6838 hornblende quartz diorite protomylonite 7125 –7128 chloritized quartz diorite protomylonite

							6838	hornblende quartz diorite protomylonite
							7125–7128	chloritized quartz diorite protomylonite

47	Legrand 1	The Ohio Oil Co.	3.0 mi SE. of Covina	1944–45	1068	6141	0 – 915±	Upper Miocene siltstone, sandstone and shale. Abundant lower Mohnian* Foraminifera
							915±–2810	Topanga formation. Sandstone, siltstone, shale and conglomerate
							915±–1220±	Buzzard Peak conglomerate member
							1220±–2810	siltstone, sandstone, and shale with Luisian* and Relizian* Foraminifera
							2810 –5800±	Glendora volcanics
							3332±	glassy basalt and holocrystalline andesite with albite and laumontite (?)
							3857 –3869	very fine-grained andesite(?) with laumontite
							4360 –4370	andesite(?)
							4895±	spilitic(?) andesite or dacite
							5646 –5651	andesite or basalt
							5800±–6082±	Mountain Meadows dacite
							6082±–6141	albite granite and granite gneiss
48	Bayly 1	Continental Oil Co.	2.1 mi W. of Spadra	1945	1092	4007	0 –2115	Topanga formation, Buzzard Peak conglomerate member
							2115±–3880±	Glendora volcanics. Largely fragmental; considerably altered, wih sulfides. Probably chiefly andesitic. Top probably interbedded with sediments
							3880±–4007	Mountain Meadows dacite
49	Walnut Creek 1	McMillan Oil Co.	1.5 mi N. of Spadra	1943	1175	2106	0 –1000±	Lower member of Puente formation
							0– 265	conglomerate lens
							265–1000±	gray sandy shale, with some Foraminifera
							1000±–1900±	Glendora volcanics
							1000±–1523	basaltic (one core)
							1523 –1900±	probably andesitic
							1900±–2106	Basement complex; chips of pink mylonitic gneiss

* Kleinpell's stage.

logically similar units at widely separated localities. Lacking evidence of the latter, and because diversity ranging from olivine basalt to rhyolite and from quiet thin flows to peléan explosions exceeds the versatility of most Recent volcanoes, it seems wise to postulate several small vents in addition. Their relative isolation and independence would explain the fact that at least nine different rock types are found in depositional contact with the basement complex within the 7.3 square miles of exposures shown on Plate 1 alone.

Most of the direct evidence for vents has already been mentioned. What is known about the dikes of the area is not suggestive, although a careful study of those in the basement complex might be rewarding. The best established of the possible exposed vents are Mystic Canyon in the foothills area (filled chiefly with pyroclastics), the dacite breccia southeast of Johnstone Peak, Elephant Hill (filled with glassy tuff and breccia), and a 60-foot possible pipe filled with perlitic breccia at the bend in Walnut Creek 0.5 mile below Puddingstone Dam. None of the material filling these vents(?) corresponds closely to the other exposed ejectamenta.

Summary: The Volcanic Episode Reconstructed

The early part of the record is out of reach in the deep central and northern parts of the Los Angeles Basin; the margins suggest that in Upper Cretaceous and Eocene time the area was receiving marine sediments, but the basin form was not so well developed as it was to become later. The seas probably withdrew during most of the Oligocene and started coming back early in the Miocene. The basin began to subside in mid-Miocene time and reached its climax of depth and localization in the upper Miocene and Pliocene.

About the time of this Miocene deepening, and perhaps genetically related to it, volcanism began on the east side of the Basin, though farther west eruptions may have started earlier in the Miocene. Toward the end of middle Miocene (Topanga) time the shore line probably passed just west of Elephant Hill (which persisted as a headland or island well into upper Miocene (middle Puente) time), through Puddingstone Reservoir, thence northward to the San Gabriel Mountain front. That these mountains were contributing to this sea is strongly suggested by the pebble suites of Topanga and Puente conglomerates (Woodford *et al.*, 1946). West of this line the highest Topanga beds are marine and contain many large boulders (in the Glendora South Hills), and the volcanics exhibit all their evidence of submarine accumulation—interbedding with fossiliferous sediments, pillow structure, palagonite, sandstone dikes, and fossiliferous tuff. East of the line the volcanics show evidence only of subaerial accumulation with very local reworking as if by streams.

In the area of present outcrops volcanism began with eruption, onto basement, of lavas and tuffs ranging from basaltic to rhyolitic, probably from several vents. At all times some of these were on the landward side of the shore line, whereas others, especially toward the end, either erupted into the sea or from offshore vents. At least one vent was emitting large volumes of tuff breccia with explosive activity that sometimes reached peléan character. The scene may have been somewhat like that near the Bay of Naples in historic time where Vesuvius and the Phlegraean Fields have carried out approximately the postulated roles.

By upper Miocene time local volcanic activity had almost completely stopped, and in the spreading Puente sea finer sediments, including a few ash falls, were deposited on the Topanga sandstones or directly on the volcanics, probably covering them all. Volcanic pebbles are more abundant in middle and upper Puente conglomerates than they are in the lower Puente, suggesting that the latter had to be removed to uncover volcanic rocks in at least part of this area.

References Cited

Alf, R. M., 1943, Mylonites in the eastern San Gabriel Mountains: Calif. Jour. Mines Geology, v. 34, p. 145–151.

—— 1948, A mylonite belt in the southeastern San Gabriel Mountains, California: Geol. Soc. America Bull., v. 59, p. 1101–1120.

Anderson, C. A., 1933, The Tuscan formation of northern California with a discussion concerning the origin of volcanic breccias: Univ. Calif. Pub., Bull. Dept. Geol. Sci., v. 23, p. 215–276.

Bacon, C. S., Jr., 1947, Applications of the Niggli-Becke projection for rock analyses: Am. Mineralogist, v. 32, p. 257–295.

Böse, E., and Wittich, E., 1913, Informe relativo a la exploración de la region norte de la costa occidental de la Baja California: Parergones del Inst. Geol. Mex., v. 4, p. 307–529.

Bramlette, M. N., 1946, The Monterey formation of California and the origin of its siliceous rocks: U. S. Geol. Survey Prof. Paper 212, 57 pages.

Bremner, C. St. J., 1932, Geology of Santa Cruz Island, Santa Barbara County, California: Santa Barbara Mus. Nat. History Occ. Papers no. 1, p. 5–33.

Daly, R. A., 1933, Igneous rocks and the depths of the earth: 2d ed., N. Y., 598 pages.

Eckis, Rollin, 1928, Alluvial fans of the Cucamonga district, southern California: Jour. Geology, v. 36, p. 224–247.

—— 1934, Geology and ground water storage capacity of valley fill: Calif. Div. Water Resources Bull. 45, 279 pages.

Elam, J. G., 1948, Geology of Seminole quadrangle, Los Angeles County, California: Univ. Calif. Los Angeles Unpub. M. A. thesis.

Eldridge, G. H., 1907, The Puente Hills oil district, southern California: U. S. Geol. Survey Bull. 309, p. 102–137.

English, W. A., 1926, Geology and oil resources of the Puente Hills region, southern California: U. S. Geol. Survey Bull. 768, 110 pages.

Fuller, R. E., 1938, Deuteric alteration controlled by the jointing of lavas: Am. Jour. Sci., 5th ser., v. 35, p. 161–171.

Geijer, P., 1913, On poikilitic intergrowths of quartz and alkali feldspar in volcanic rocks: Geologiska Foreningens, Forhandlingar, Stockholm, Bank 35, p. 51–80.

Hinds, N. E. A., 1934, Jurassic age of last granitoid intrusives in the Klamath Mountains and Sierra Nevada, California: Am. Jour. Sci., 5th ser., v. 27, p. 182–192.

Hoots, H. W., 1931, Geology of the eastern part of the Santa Monica Mountains, Los Angeles County, California: U. S. Geol. Survey Prof. Paper 165-C, 134 pages.

Iddings, J. P., 1899, *in* Hague, Arnold, Iddings, *et al.*, Geology of the Yellowstone National Park: U. S. Geol. Survey Mon. 32, pt. II, 893 pages.

Kleinpell, R. M., 1938, Miocene stratigraphy of California: Am. Assoc. Petroleum Geologists, Tulsa, 450 pages.

La Croix, 1904, La Montagne Pelée et ses eruptions: L'Acad. des Sci., Paris, 662 pages.

—— 1908, La Montagne Pelée apres ses eruptions avec observations sur les eruptions du Vesuve en 79 et en 1906: L'Acad. des Sci., Paris, 136 pages.

Larsen, E. S., Jr., 1948, Batholith and associated rocks of Corona, Elsinore, and San Luis Rey quadrangles, California: Geol. Soc. America Mem. 29, 182 pages.

MacGregor, A. G., 1938, The Royal Society expedition to Montserrat, B. W. I. The volcanic history and petrology of Montserrat, with observations on Mt. Pelée, in Martinique: Royal Soc. London Philos. Trans., ser. B, no. 557, v. 229, 90 pages.

Mendenhall, W. C., 1908, Ground waters and irrigation enterprises in the Foothill Belt, southern California: U. S. Geol. Survey W.-S. Paper 219, 180 pages.

Miller, W. J., 1934, Geology of the western San Gabriel Mountains of California: Pub. Univ. Calif. Los Angeles in Math. and Phys. Sci., v. 1, no. 1, 114 pages.

Mull, B. H., 1934, Miocene volcanics of the San Jose Hills: Univ. Calif. Berkeley, Dept. Geology, Unpub. M. A. thesis.

Neumann van Padang, M., 1934a, Die Eruptionsregenfrage in Bezug auf den grossen Krakatau-Ausbruch vom 26 und 27 August 1883: Koninklijke Akad. Van Wetenschappen te Amsterdam Proc., v. 37, no. 3, p. 1–8, 168–173.

—— 1934b, Haben bei den Ausbruchen des Slametvulkans Eruptionsregen stattgefunden?: Leidsche Geol. Mededeelingen, Deel VI, Blz. 79–97.

Norton, W. H., 1917, A classification of breccias: Jour. Geology, v. 25, p. 160–194.

Oakeshott, G. B., 1938, Geology and mineral deposits of the western San Gabriel Mountains, Los Angeles County: Calif. Jour. Mines Geology, v. 33, p. 216–249.

Perret, F. A., 1935, The eruption of Mt. Pelée 1929–1932: Carnegie Inst. Pub. 458.

Popenoe, W. P., 1942, Upper Cretaceous formations and faunas of southern California: Am. Assoc. Petroleum Geologists Bull., v. 26, p. 162–187.

Robinson, H. H., 1913, The San Franciscan volcanic field, Arizona: U. S. Geol. Survey Prof. Paper 76, 213 pages.

Schoellhamer, J. E., Kinney, D. M., Yerkes, R. F., and Vedder, J. G., 1954, Geologic map of the northern Santa Ana Mountains, Orange and Riverside counties, California: U. S. Geol. Survey Oil and Gas Inv., Map OM 154.

Schoellhamer, J. E., and Woodford, A. O., 1951, The floor of the Los Angeles Basin: U. S. Geol. Survey Oil Map 117.

Shelton, J. S., 1946, Geology of northeast margin of San Gabriel Basin, Los Angeles County, California: U. S. Geol. Survey Oil and Gas Inv. Prelim. Map 63.

—— 1954, Miocene volcanism in coastal southern California: Calif. Div. Mines Bull. 170,

Stone, Benton, 1950, *"Lagena samanica"* Berry: The Micropaleontologist, v. IV, no. 2, p. 17.

Taliaferro, N. L., Hudson, F. S., and Craddock, W. N., 1924, Oil fields of Ventura County, California: Am. Assoc. Petroleum Geologists Bull., v. 8, p. 789–829.

Tröger, W. E., 1935, Spezielle Petrographie der Eruptivgesteine: Berlin, 360 pages.

Wentworth, C. K., and Williams, H., 1932, The classification and terminology of the pyroclastic rocks: Natl. Research Council Bull. 89, p. 19–53.

Williams, H., Turner, F. J., and Gilbert, C. M., 1954, Petrography: San Francisco, 406 pages.

Winchell, A. N., 1951, Elements of optical mineralogy, Part II: N. Y.,

Woodford, A. O., 1939, Pre-Tertiary diastrophism and plutonism in southern California and Baja California: 6th Pacific Sci. Cong. Proc., p. 253–258.

Woodford, A. O., Shelton, J. S., and Moran, T. G., 1944, Stratigraphy and oil possibilities of Puente and San Jose Hills, California: U. S. Geol. Survey Oil and Gas Inv., Prelim. Map 23.

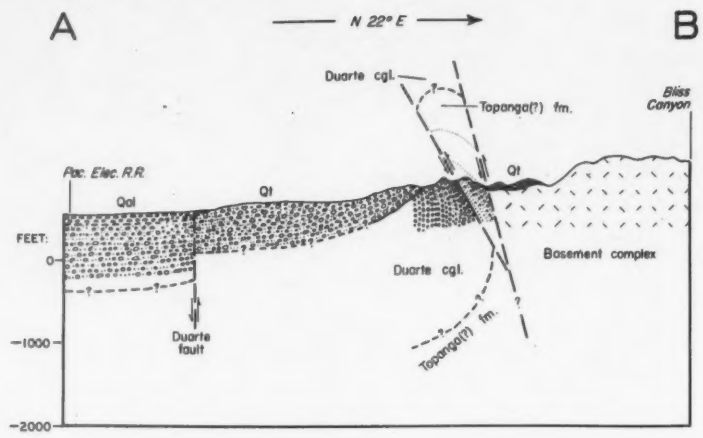
—— 1946, Miocene conglomerates of Puente and San Jose Hills, California: Am. Assoc. Petroleum Geologists Bull., v. 30, p. 514–560.

Woodring, W. P., Bramlette, M. N., and Kew, W. S. W., 1946, Geology and paleontology of Palos Verdes Hills, California: U. S. Geol. Survey Prof. Paper 207, 145 pages.

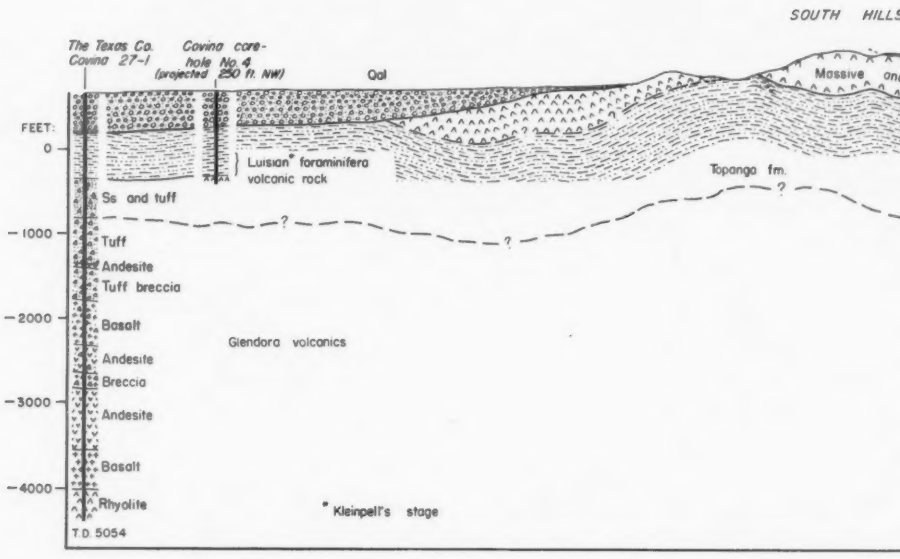
Pomona College, Claremont, California

Manuscript Received by the Secretary of the Society, August 1, 1951

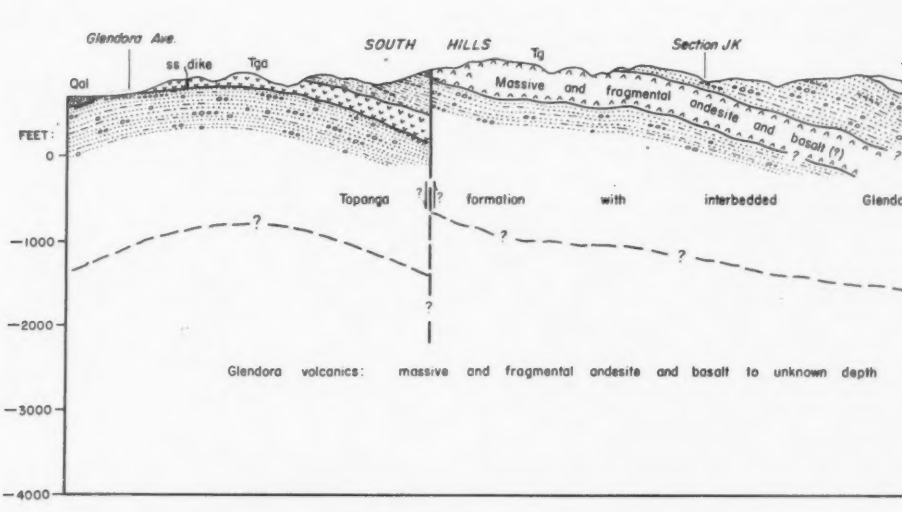
Project Grant 461-45


A
N 22° E
B
Duarte cgl.
Topanga(?) fm.
Bliss Canyon
Pac. Elec. R.R.
Qal
Qt
Qt
FEET:
0
-1000
-2000
Duarte fault
Duarte cgl.
Basement complex
Topanga(?) fm.

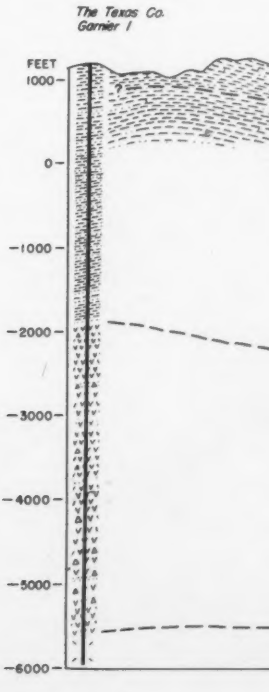
J

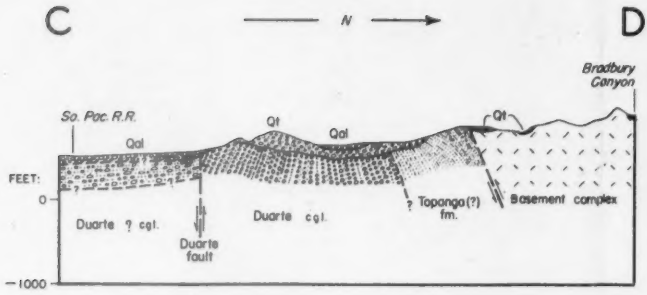

SOUTH HILLS
The Texas Co. Covina 27-1
Covina core-hole No. 4 (projected 250 ft. NW)
Qal
Massive
FEET:
0
-1000
-2000
-3000
-4000
Luisian* foraminifera volcanic rock
Topanga fm.
Ss and tuff
Tuff
Andesite
Tuff breccia
Basalt
Andesite
Breccia
Andesite
Basalt
Rhyolite
T.D. 5054
Glendora volcanics
* Kleinpell's stage

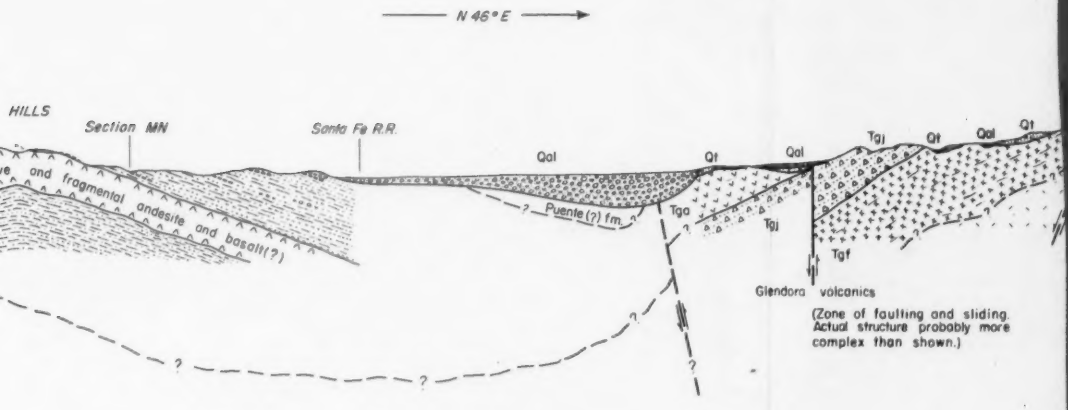
M

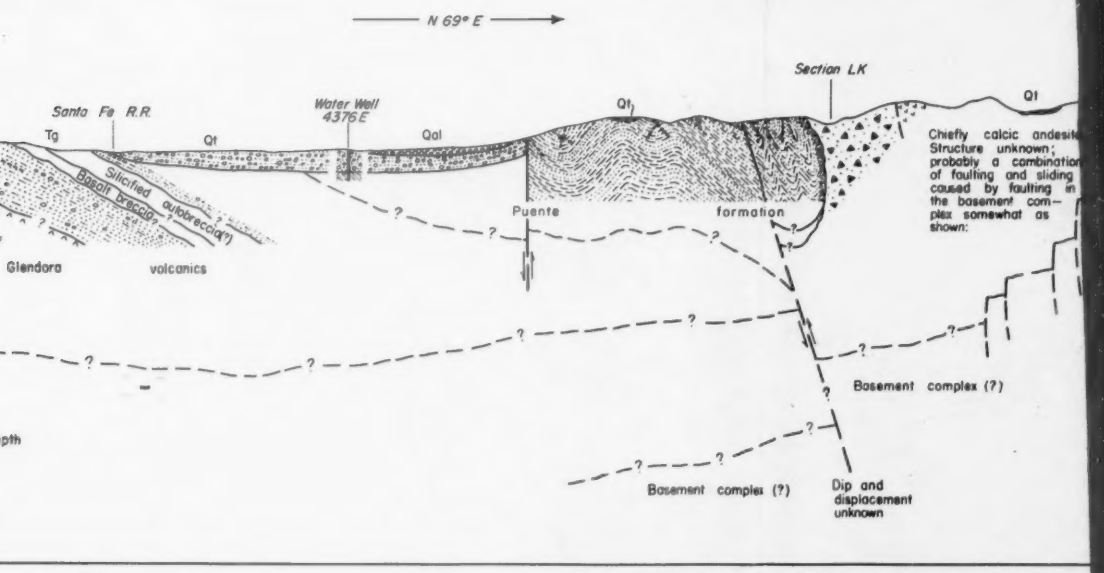

Glendora Ave.
SOUTH HILLS
Section JK
ss dike
Tga
Tg
Qal
Massive and fragmental andesite and basalt (?)
FEET:
0
-1000
-2000
-3000
-4000
Topanga formation with interbedded
Glendo
Glendora volcanics: massive and fragmental andesite and basalt to unknown depth

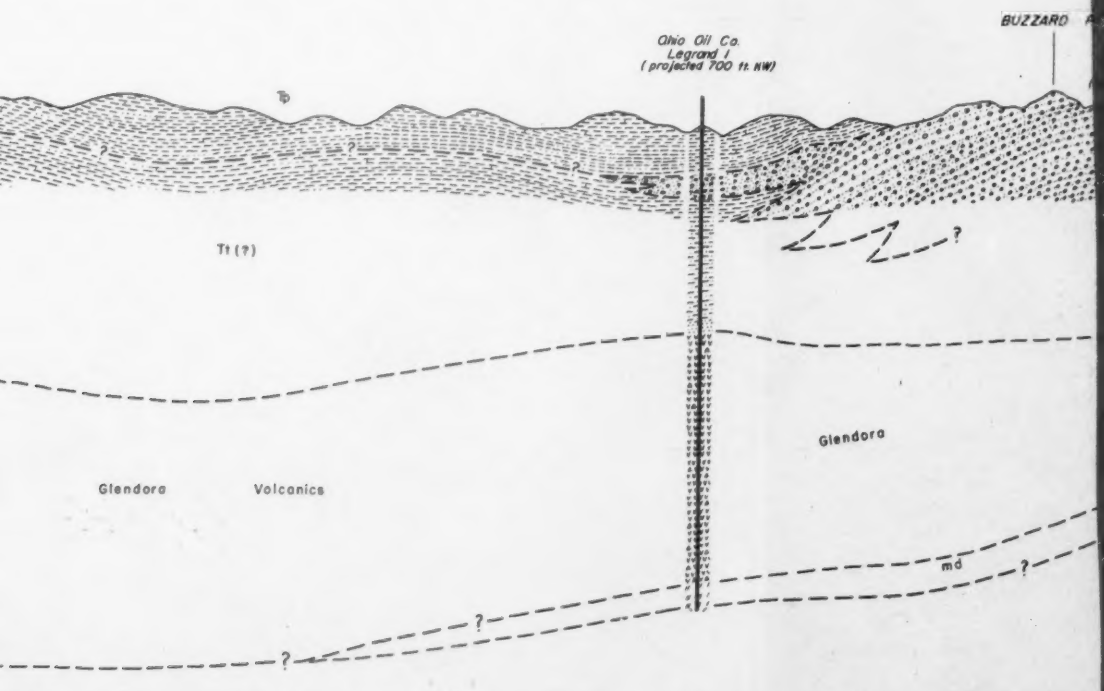
R


The Texas Co. Garnier 1
FEET
1000
0
-1000
-2000
-3000
-4000
-5000
-6000



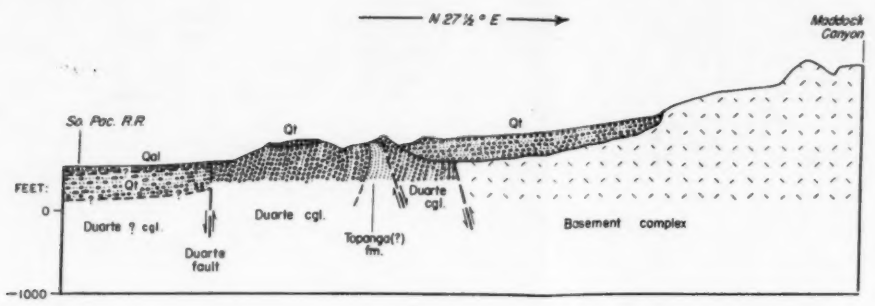






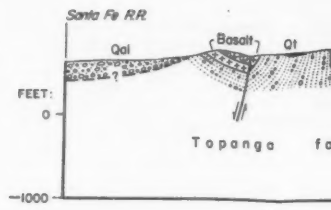
GEOLOGIC STRUCTUR

C

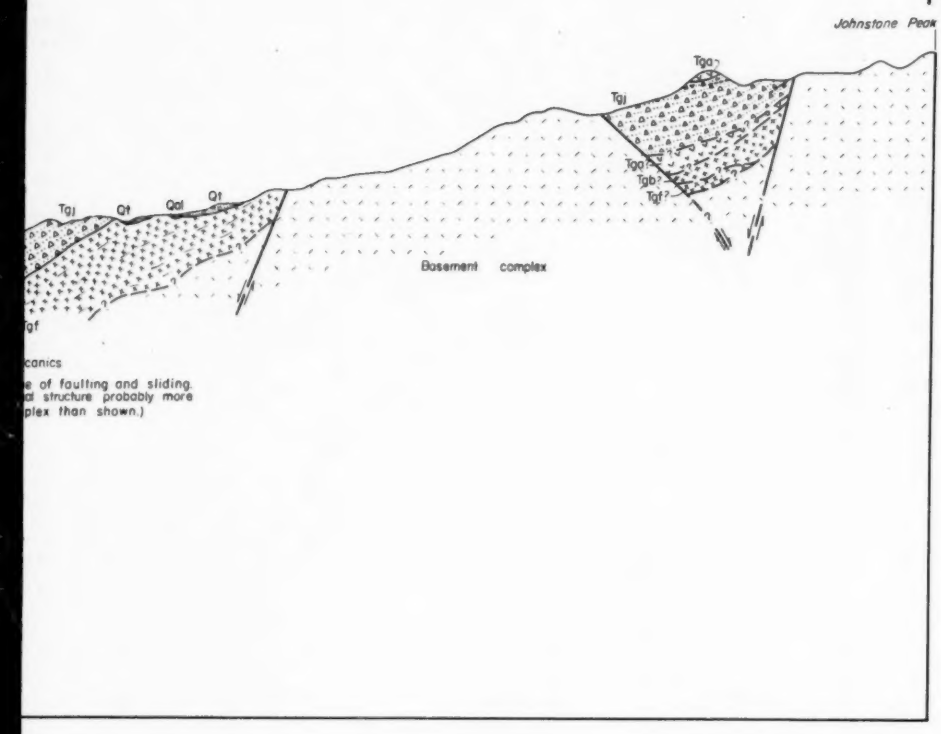


E

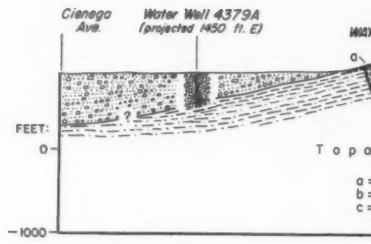
F



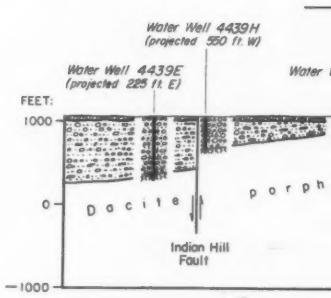
K



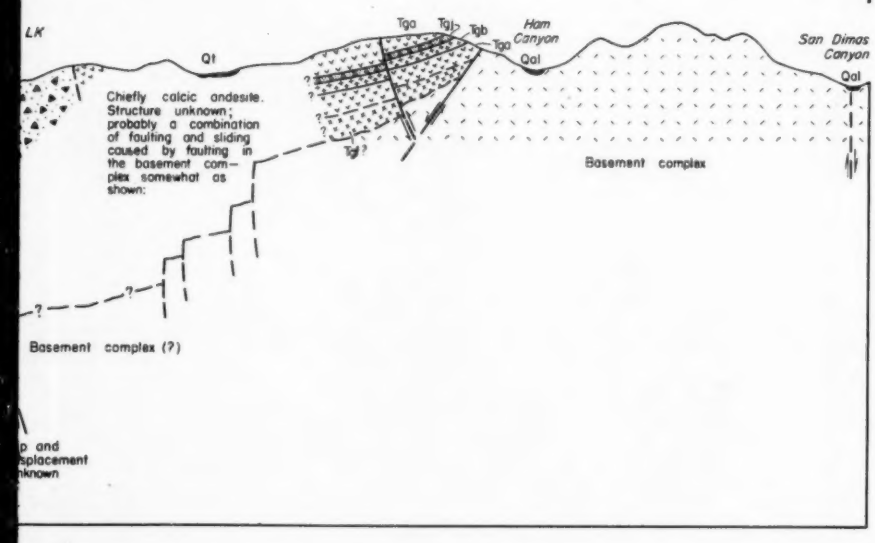
L



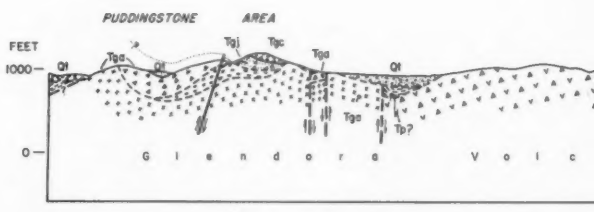
O



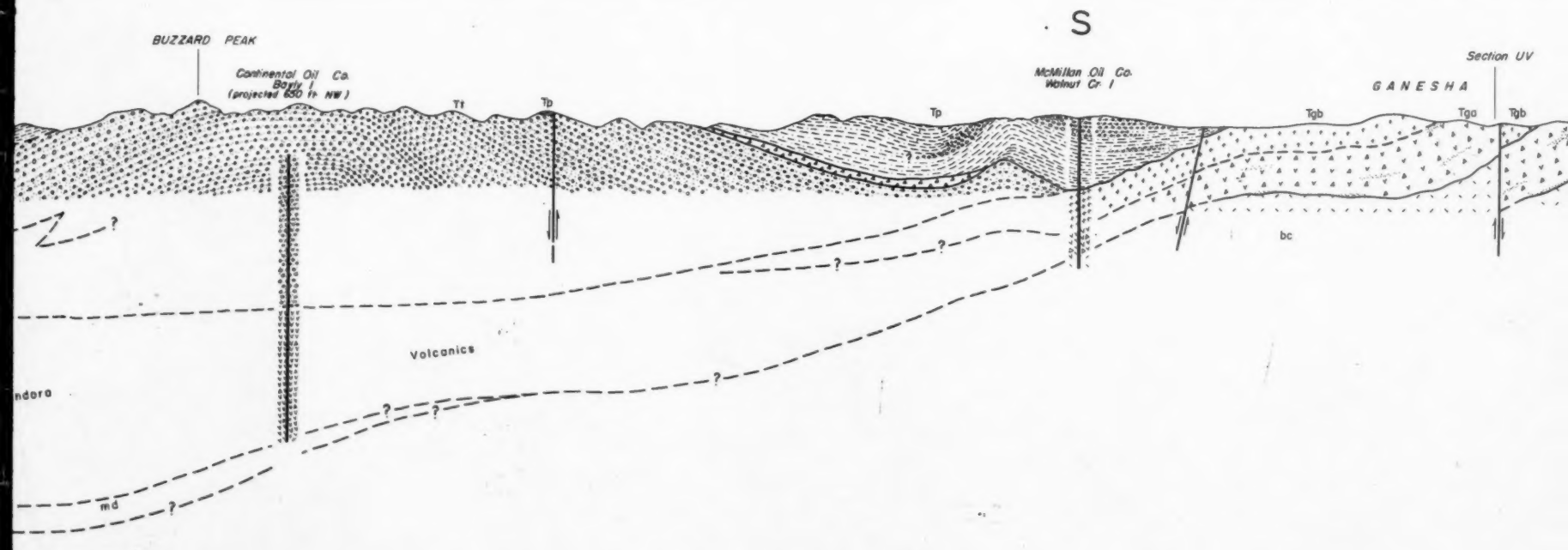
N



U



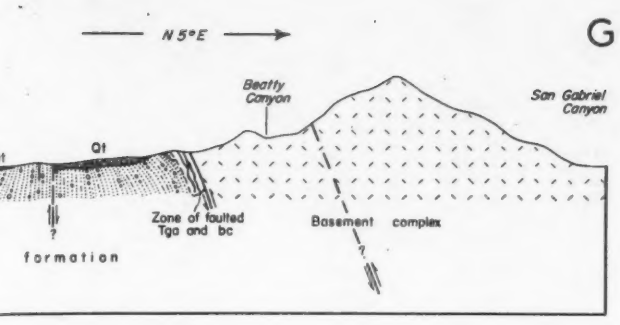
S

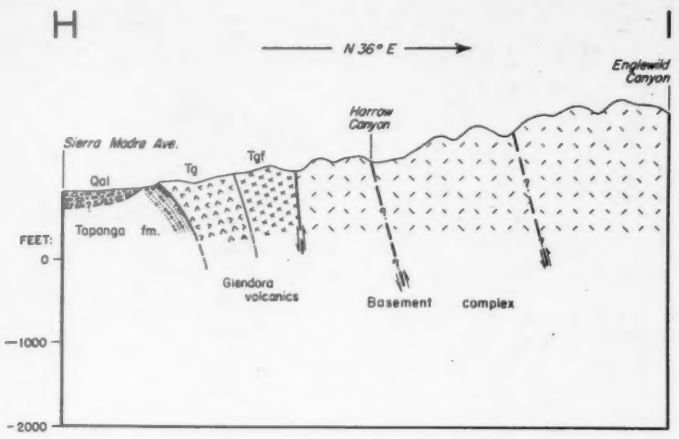


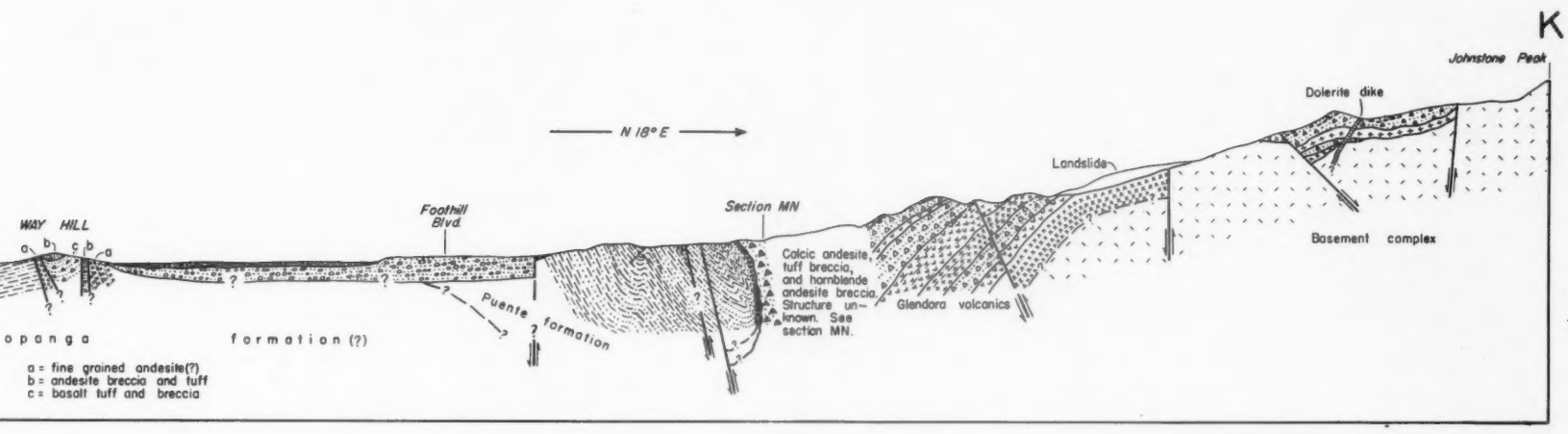
OGIC STRUCTURE SECTIONS, GLENDORA VOLCANIC ROCKS AND DUARTE CONGLOMERATE, CALIFORNIA

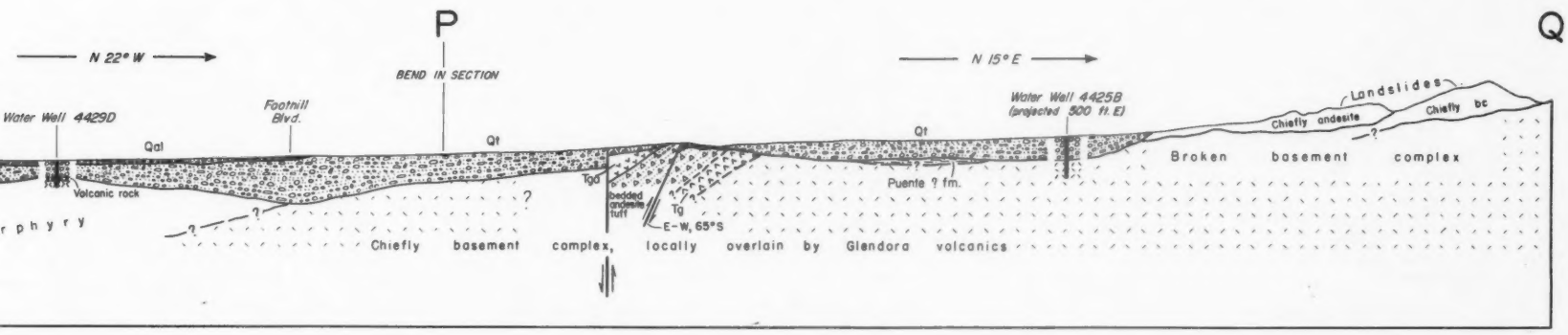
Scale 1:24000

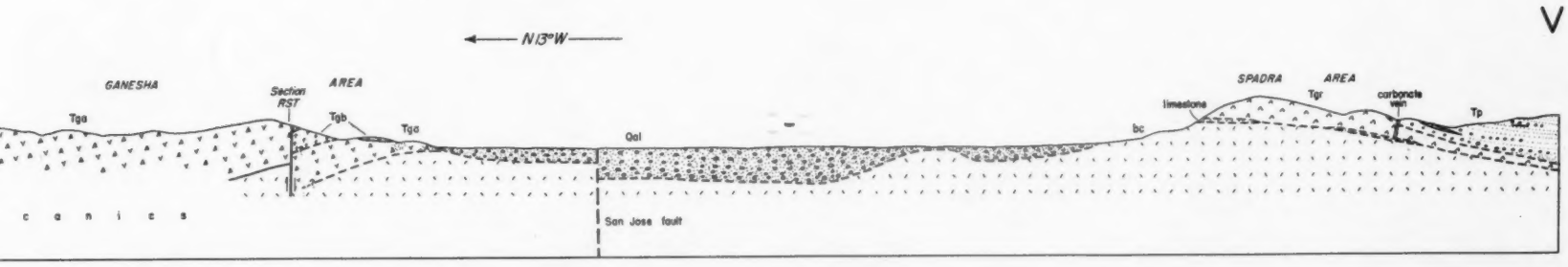
See Plate 1 for locations and explanation of symbols

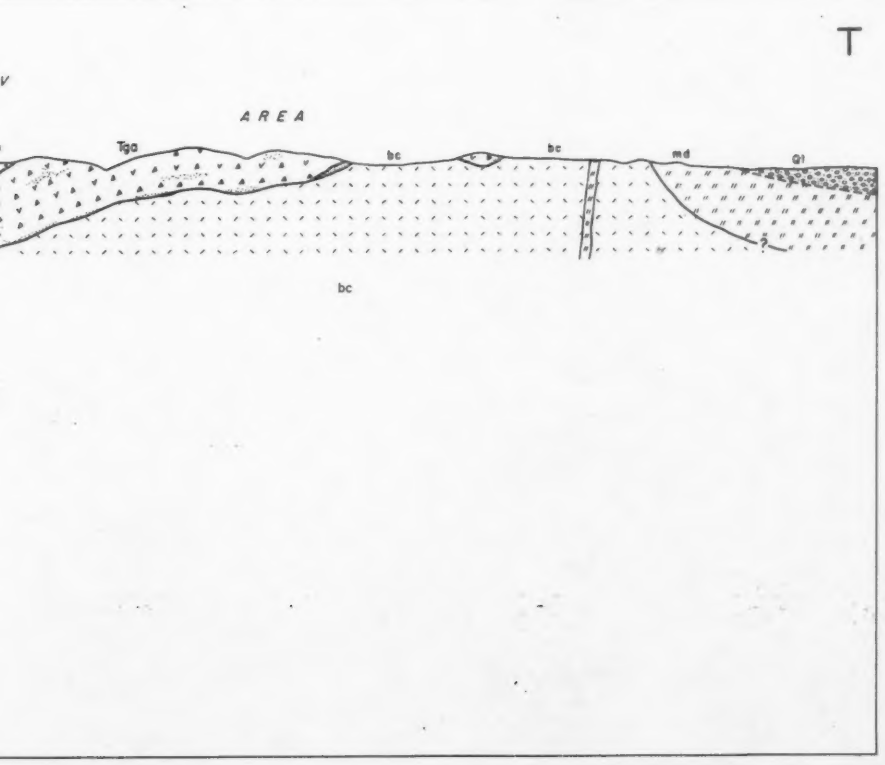

N 5°E
G
Beatty Canyon
San Gabriel Canyon
Qt
Zone of faulted Tga and bc
Basement complex
formation


H
N 36°E
I
Englewild Canyon
Harrow Canyon
Sierra Madre Ave.
Qal
Tg
Tgf
Tapanga fm.
FEET:
0
-1000
-2000
Glendora volcanics
Basement complex


K
Johnstone Peak
Dolerite dike
N 18°E
Landslide
WAY HILL
Foothill Blvd.
Section MN
Basement complex
Calcic andesite, tuff breccia, and hornblende andesite breccia. Structure unknown. See section MN.
Glendora volcanics
Puente formation
opanga
formation (?)
a = fine grained andesite(?)
b = andesite breccia and tuff
c = basalt tuff and breccia


P
N 22°W
BEND IN SECTION
Water Well 4429D
Foothill Blvd.
Qal
Qt
Volcanic rock
Tgd
bedded andesite tuff
Tg
E-W, 65°S
Chiefly basement complex, locally overlain by Glendora volcanics
Q
N 15°E
Water Well 4425B (projected 500 ft. E)
Landslides
Chiefly andesite
Chiefly bc
Broken basement complex
Puente ? fm.


V
N 13°W
GANESHA
AREA
Section RST
Tga
Tgb
Qal
bc
SPADRA
AREA
limestone
Tgr
carbonate vein
Tp
San Jose fault


T
AREA
Tga
bc
md
Ql

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 66, PP. 91-104, 1 FIG., 4 PLS. JANUARY 1955

PERIDOTITES OF MANPUR, SINGHBHUM DISTRICT, BIHAR, INDIA, AND ORIGIN OF ASSOCIATED ASBESTOS DEPOSITS

By S. C. Chatterjee

Abstract

An occurrence of a workable deposit of tremolite asbestos has originated by the alteration of a body of ultrabasic rock intrusive into the Archean crystalline rocks of the Singhbhum District, Bihar, India. Although there is microscopic evidence of the formation of antigorite or chrysotile from olivine, and tremolite from diallage, lime metasomatism has converted the chrysotile into logs of tremolite asbestos. Chrysotile asbestos also occurs, but it is subordinate to tremolite. Serpentinization was caused by hydrothermal solutions derived from some unknown deep source since younger acid intrusives are lacking in the area. Olivine grains have been affected uniformly, and serpentinization is not confined to their cores or margins, nor to any pre-existing fissures. The nonfibrous serpentine was transformed into fibrous asbestos by stress acting along certain zones. The ultrabasic bodies are not inclusions in the granite gneiss but represent younger intrusions.

CONTENTS

TEXT

	Page
Introduction	91
Acknowledgment	92
Geological setting	92
Granitic rocks	94
Dioritic rocks	95
Epidioritic types	96
Peridotites	96
Chemical composition of younger ultrabasic and of tremolite asbestos	101
Origin of asbestos	101
Age of intrusive, its tectonic setting, and relation to other ultrabasic occurrences in Singhbhum	103
Conclusion	104
References cited	104

ILLUSTRATIONS

Figure	Page
1. Geological map of Manpur area	92

Plate	Following page
1. Asbestos quarry and asbestos bearing peridotite	98
2. Microsections of diorite and peridotites showing various stages in the formation of asbestos veins	98
3. Microsections showing the formation of secondary minerals	98
4. Microsections showing the transformation of serpentine to asbestos, replacement of chrysotile, and effects of shearing	98

Introduction

The only published geological account of the Manpur (Fig. 1) area is that given by Dunn and Dey (1942) in their Memoirs on the Geology of East Singhbhum. The peridotite rocks were not described by them, presumably because no fresh rock was found near the surface. They have, however, given a good petrographical description of the rocks in which the ultrabasics occur. The basement rock, which is mainly dioritic, was intruded into Dharwar quartzites and phyllites. Outcrops of quartzite occur in the northwest where they form the hillock near Dabanki and in the south of the area studied, with intercalations of phyllites. Inclusions of hornblende schist, and small inclusions of phyllite in the diorite, were noted by Dunn and Dey (p. 398). The lighter granitic bands in the diorite were considered intrusive into the diorite. The transformation of the diorite into chlorite and talc schists by shearing was also noted by these authors (p. 398). The authors assumed, however, that the ultrabasic rock is an inclusion in the diorite, and they observed "An inclusion of ultrabasic rock south of Manpur (22°36′:86°16′) contains asbestos" (p. 398). The present writer believes that the diorite is not a true diorite of igneous origin but is a metamorphic hornblende-plagioclase-quartz rock of low-grade epidiorite type, and that the ultrabasic bodies are not inclusions but younger intrusions. The term diorite will, how-

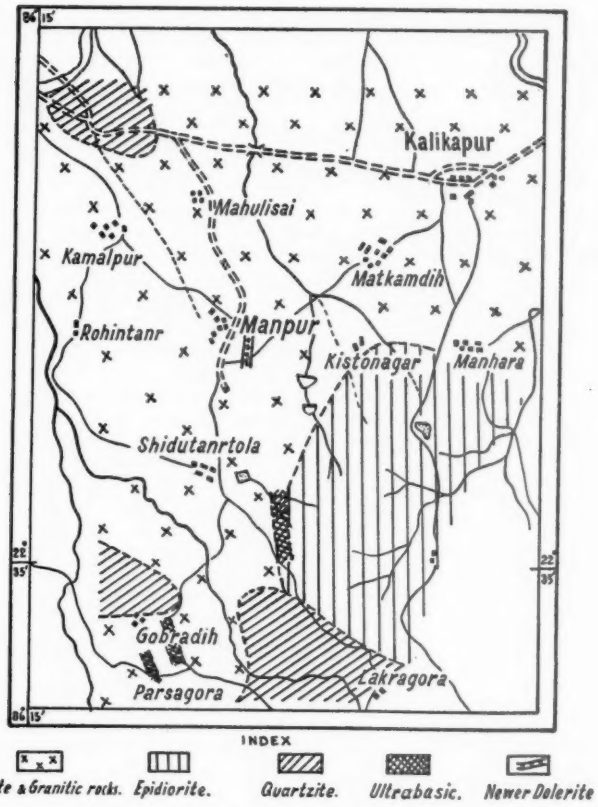


FIGURE 1.—GEOLOGICAL MAP OF MANPUR AREA
Scale 1" = 1 mile

ever, be retained as a field term to avoid confusion.

ACKNOWLEDGMENT

The author wishes to acknowledge his gratefulness to M. G. Rungta and Sons, the owners of the asbestos mines, for kindly permitting study of these mines, to Mr. Umapada Chatterjee, Zamindar of the Manpur area, for providing facilities for work, to Mr. W. H. Herdsman of Glasgow for kindly undertaking the chemical analysis of the ultrabasic rock and a sample of asbestos.

GEOLOGICAL SETTING

The dioritic rocks are well jointed, foliated, and sheared in certain zones; the foliation parallels the major joints. The principal joints run northwest-southeast, cutting the rock into bands and ridges which also trend northwest-southeast, and dip steeply northeast. Composition varies from granite to gabbro, although the bulk of the rock is a banded diorite in which the granite occurs as bands. There are also veins of quartz. In many places the bands are bent and even contorted. Southeast of Manpur village the diorite is dark, with light bands, and farther south it passes into a hornblende-rich type. Toward the west it becomes gabbroic. The granitic rocks (Singhbhum granite) to the north and northeast are younger than the diorite; dikes of pegmatitic granite cut the gabbroic rocks.

The banded diorite is not only contorted and

bent, but in certain localities is much sheared with resulting slickensided surfaces and a talcose appearance, representing zones of faulting and crushing with a low-grade mineral assemblage of epidote, chlorite, sericite, and talc. The diorite has thus had a complicated metamorphic history; the rock has been permeated by granitizing liquids in many places and later subjected to low-grade metamorphism in certain crush zones. In fact the so-called diorites are due to the amphibolitization of the pyroxenes of pre-existing basic rocks and actually belong to the albite epidote-amphibolite facies; with increase of lighter constituents they resemble granitic rocks.

Apart from the ultrabasic intrusives, there are in the diorite inclusions of another much-altered ultrabasic rock and of a basic rock; these range from thin discontinuous bands to dikelike masses. Southeast of Manpur and east of the main diorite hillocks stands a large hillock (Pararakha *dungri*) consisting of a dark ultrabasic rock. This rock forms dikelike masses trending parallel to the principal joint slabs of diorite (NW.–SE.). The ultrabasic rock is also cut into large, thin slabs by joints. Microscopic examination shows this rock to be metamorphosed and to resemble the type described as "epidiorite" by Dunn and Dey (1942, p. 367). In places it is traversed by veins and irregular patches of quartz; in others it contains chalcopyrite.

The ultrabasic bodies do not form dikes on the surface but a jumbled mass of boulders in all states of alteration. They weather easily. Fresh rocks are found in pits dug for mining asbestos where there is a rapid and irregular variation from hard, black, and greenish-black peridotite to talcose and talc-chlorite-tremolite rocks, through various stages of serpentinization, steatitization, and chloritization.

In the earlier stages of mining the pits were isolated; later they were joined end to end, so that now two large bodies are exposed south of Manpur and one near Gobradih. The Manpur mine now extends about a mile, and the larger body is about 50 feet deep. At Gobradih there are only surface quarries.

Fresh rocks were obtained from the Manpur pit southeast of the large tank. Here black and greenish-black, massive peridotite appears to expand downward under the dioritic rocks, outcrops of which occur east and northeast of the peridotite mass. The exact junction is hidden under thick tropical soil and vegetation. The ultrabasic is much jointed; the principal joints strike southeast—northwest to north-south and have produced ridges. This direction conforms to the strike of the diorite joint slabs. The joint planes are almost vertical or dip steeply. A second set of vertical joints trend east-west and east-southeast–west-northwest almost at right angles to the first vertical set.

Two sets of oblique joints strike east-west; the first set dips 45°–60°N., and the second set 70°N. In the larger, more southern, of the two pits at Manpur, the ultrabasic is much jointed, and there are three sets of vertical joints, the most prominent of which trends east-west or east-northeast–west-southwest. In addition, two oblique joints dip in opposite directions—southwest and northeast.

The pit is cut irregularly, following the trend of workable veins with branches from the main trench. The thickness of the ultrabasic is undetermined. At one place a contact between the ultrabasic and dioritic granite is exposed, but there was no evidence of contact metamorphism.

The most prominent east-west joints are wide and are cross joints or tension joints; the other vertical set are longitudinal joints. The rocks are traversed by minor joints, which are irregular and are perceptible only when the rock is broken. The asbestos veins penetrate these joints. The smaller veins pass along the minor joints and intrude the rock between joints (Pl. 1, fig. 1). Starting from the joint planes, serpentinization extends into the mass of the rock and, in the last stage, replaces the entire mass by asbestos veins.

The parent ultrabasic, when fresh, is uniform, but the products of alteration are not the same everywhere. Owing to unsystematic mining, no chronology for the secondary minerals could be found. In many places the dark, apparently unaltered rocks show veins of cross-fiber chrysotile extending on either side of a median fracture with narrow islets of unreplaced rock in between. Only rarely do the fibers consist of pure chrysotile; generally they are harsh and tremolitic. Microscopic examination shows that the fibers are composite and mainly tremolitic

(Pl. 1, fig. 2). Elsewhere similar fibers consist of tremolite; the contact between the serpentinized rock and these fibers is irregular, and there is no median fracture. Gradations range from greenish woody actinolitic fibers to white tremolite. Actinolite and chlorite color the rock green.

A third mode of occurrence is filling of the joints or fractures. Precipitation resulted in asbestos fibers within the joint planes. The asbestos is also tremolitic in these veins. The junction between the asbestos and the rock is sharp so that the asbestos can be scraped off. The rock is compact and granular, but, whereas in the first two cases it is serpentinized, here it is not so highly altered.

The ultrabasic is also traversed by irregular zones of talc schist, talc-chlorite schist, and tremolite-calcite rocks which are not only highly sheared but also crumpled, corrugated, and drag-folded. These zones of altered rock represent original shear zones and zones of fracture which facilitated the movements of the hydrothermal fluids. The shearing movement continued even after the tremolitization and steatitization of the rocks. Microscopic evidence shows that (1) the fibrous rocks have a parallel arrangement of the magnetite granules separated during serpentinization, and (2) the talc has been drawn out into fine shreds and streaks with relics of anthophyllite. The texture resembles that produced by fluxional movement. In the serpentine bodies of Unst in the Shetland Islands, Read (1934, p. 666) has shown that the talc-magnesite-chlorite rocks, associated with a bottom layer of antigorite, are related to definite tectonic lines and are the result of stress metamorphism. The bottom antigoritic layer was produced during upward movement of the ultrabasic body. Talc schists were formed at the sole, and incipient thrusting gave rise to talcose layers striking parallel to the main thrust front and to similar talcose belts due to tears at right angles to the first. More recently Amin (1952, p. 97) has shown "that tectonic deformation was accompanied by the passage of hydrothermal fluids along the movement-planes which was responsible for the transformation of the serpentines", but while localized shearing has caused metamorphic differentiation of the components into pure masses of talc, chlorite, and carbonates, in the present area there has been no such segregation of the individual components.

Granitic Rocks

The typical granite is light pink. Microscopic examination shows that the rock is gneissic; granulated quartz alternates with feldspar in subparallel rows. The feldspar is of three types —microcline with incipient cross-hatching, untwinned orthoclase with straight extinction, and perthite and plagioclase which appears to be not more calcic than oligoclase. Perthite is abundant. The feldspars are partially clouded. Small flakes and sheafs of biotite also occur as was noted by Dunn and Dey (1942, p. 399). The accessories are small granules of zircon and needles of apatite.

The banded granitic gneiss near Rajbandh shows alternate quartz and feldspar bands in the white portion. Some of the feldspar is orthoclase; the rest is plagioclase in different stages of alteration. The small quantity of unaltered plagioclase is albitic. Twinned feldspars are not oriented in the direction of banding as shown by the angular direction of the twin lamellae. Quartz is granulated with undulatory extinction. Abundant sphene, in lozenge-shaped crystals and granules, and epidote have formed at the expense of the feldspar. The dark minerals are hornblende with ragged edges, and chlorite. The hornblende has the same characteristics found in the hornblende of the other groups.

Near the contact with the ultrabasic, the granitic rock shows alternating thin bands of finely granulated mosaic of quartz and feldspar and largely feldspathic bands, in which the porphyroclasts of feldspars have crenulated edges with their small bays filled with minute granules. Some of the grains have clear contacts with each other, while the other sides have crenulations. Quartz has been completely granulated so that the larger remnants are almost entirely feldspars. These consist of plagioclase of albite type with broad lamellae and microperthite. One or two untwinned feldspar grains also occur. Some of the microperthites have been partially replaced by plagioclase near the edges. Most grains have wavy extinction, a few have a radial extinction.

Bunches of tremolite needles form a thick felt interspersed with granules of quartz and feldspar. The needles insinuate into the mosaic, sometimes enclosing the granules. The tremolite felts roughly parallel the banding. Other minerals present are chlorite, zircon, and magnetite.

Dioritic Rocks

Dioritic rocks are found only near Manpur village. The rocks are generally gneissoid. There are also thin bands of quartz. Under the microscope the rocks appear highly altered. Apart from mineralogical changes there are textural changes; the texture does not conform to an igneous texture but resembles that of an epidiorite, with an assemblage characteristic of the albite-epidote-amphibolite facies. Quartz is highly granulated; the granules occur as streaks showing wavy extinction and ranging from medium to very fine granules. Some of the finest granules are in micropegmatitic intergrowth. Feldspars are largely saussuritized or clouded. A few unaltered ones show albite and Carlsbad twinning; here the maximum refractive index indicates albite. In many feldspars the core is cloudy, while the outer portion is saussuritized. In some slides the feldspars show granulation. The common alteration products are sphene, epidote, allanite, and zoisite; these have drawn materials from the other minerals also.

The hornblende is strongly pleochroic with X = yellow, Y = brownish green, Z = bluish green. The last is so blue as to suggest that it is a soda-rich variety. The general habit is prismatic, but the minerals appear to have been bent and twisted in the direction of elongation imparting a "crystallization schistosity" parallel to the direction of the gneissic texture. Although a linear arrangement is dominant, random orientation is also seen. The edges are irregular and tattered with needlelike extensions and projections. The needles are actinolitic but maintain optical continuity with the hornblende. Needles of tremolite are also scattered at random in the feldspathic base and, in places, form irregular patches of hornblende. A few hornblende individuals are twinned, and some are sieved with quartz droplets. Others have droplets, veinlets, and canals of quartz. Even symplectitic intergrowths of quartz and hornblende are seen. There is, however, no trace of original pyroxene. It is clear that the needles of actinolite growing in the feldspathic base have drawn a part of their material from the feldspar. Lime liberated in the process and silica must have converted a part of the ilmenite to sphene. A stage in the reaction is marked by the coat of leucoxene around ilmenite grains. The remainder of the lime with alumina has gone into zoisite and epidote. The reactions are suggested by the intimate association of feldspar with amphibole, zoisite with amphibole, sphene with feldspar and ilmenite. There are inclusions of sphene in hornblende (Pl. 2, fig. 1). Part of the anorthite molecules also enters into chlorite which in many places is associated with the hornblende.

The hornblende in its turn shows transformation into chlorite with a sharp boundary against it in many slides, but gradual in others, easily recognized by characteristic birefringence of chlorite with optical continuity. Epidote and zoisite are also associated with the hornblende in a manner which suggests that they have formed at the expense of hornblende.

The transformation is seen clearly in the banded gneisses near the tank south of the Manpur dispensary, in which the white bands are quartzo-feldspathic, and the dark bands are greenish from the extensive development of epidote. Microscopic sections of the dark-greenish bands indicate the order of formation of the secondary minerals. One section shows that the hornblende has vague boundaries and is patchy, the patch passing into chlorite. Other hornblende individuals are breaking up into tremolite needles. Another section shows no hornblende but has a large amount of chlorite, epidote, allanite with cracks, and an almost isotropic epidote with a narrow ring of allanite. The chlorite is penninite. Sphene, pleochroic from colorless to pinkish, is abundant. The other minerals are granulated quartz and saussuritized feldspar. The rocks seem to have undergone retrogressive changes after the formation of hornblende. In the final stage epidote and chlorite have replaced the hornblende.

The grade of metamorphism therefore appears to be lower than that for the epidote-amphibolite facies. When actinolite appears

with chlorite, the assemblage actinolite + chlorite (aluminous) + epidote + albite + sphene must have arisen from augite or hornblende + plagioclase + ilmenite + water as indicated by Turner (1948, p. 52).

The rocks contain, in addition to the above minerals, an appreciable quantity of rounded inclusions of apatite in hornblende and laths of apatite.

Epidioritic Types

The rocks in the southern part of the area are more basic and resemble more the epidioritic types of Dunn (1929, p. 77). They are much more altered than the dioritic types and contain a large amount of tremolite, chlorite, and talc formed from amphiboles whose relics are recognized by their rhombic cleavage and pleochroism. The chlorite is pleochroic from yellow to green and has an isotropic interior. The feldspars are completely saussuritized. The small quantity of quartz and the feldspars are much granulated with a tendency toward micropegmatitic intergrowth. The quartz granules lack strain shadows probably owing to recrystallization. The rocks are crushed and sheared. In comparatively less altered specimens, some grains of the plagioclase feldspar are torn into two or three parts; the torn parts are relatively displaced, and the gaps filled by fine quartz granules. Some specimens are banded with quartz bands alternating with clouded feldspar bands. The quartz bands consist of bent strings of granules of quartz. Hornblende is also drawn out into strings.

In the epidioritic types near Bhalukgarha around the southern end of the ultrabasic, the rocks consist of altered hornblende, tremolite needles, saussuritized feldspar, granules of quartz, chlorite, epidote, and ilmenite surrounded by leucoxene. The microscopic characteristics are similar to those of the epidiorite described by Dunn (1929, p. 84).

The sheared varieties resemble talc-chlorite schists. Both quartz and feldspars are crushed into fine granules, and relics of plagioclase with albite twin and Carlsbad twin are very rare. The granulated quartzo-feldspathic mosaic tends toward incipient secondary micropegmatitic intergrowth. The secondary minerals are larger and consist mainly of chlorite, colorless talc with well-defined cleavage, epidote with irregular outlines, granules of zoisite, needles of tremolite, ilmenite set in patches of leucoxene, and sphene.

The ultrabasic types occur as dikelike masses inside the granitic or dioritic types in the eastern part of the area; they contain a small quantity of granulated quartz and feldspar. The chief component is an altered hornblende. It has a brownish tint and is much chloritized and steatitized. Needles of pale-green actinolite, and ilmenite with leucoxene are also present.

Peridotites

The least-altered specimens of the peridotites are dark greenish black or black, medium-grained, with a specific gravity of 2.75 to 2.90. With progressive alteration, the rocks lose their hardness, color, and compactness, passing into a greenish talc schist, a talc-chlorite schist, or a talc-chlorite-actinolite schist, and on through soft, greenish serpentinized and tremolitized varieties. The distribution of the various types is haphazard, although the highly altered types form zones or belts enclosed by the compact types, as has been noted earlier.

The primary minerals are olivine, anthophyllite, diallage, enstatite, magnetite, and chromite in order of decreasing abundance; the secondary minerals are chrysotile, tremolite, actinolite, chlorite, talc, and chalcopyrite.

Most of the olivine is altered. The unaltered olivine is traversed by cracks filled with magnetite granules. Many grains are enclosed partially by anthophyllite which may be in direct contact or with a narrow, altered zone (Pl. 2, fig. 2). Olivine is more susceptible to alteration than anthophyllite, and commonly a pseudomorph of greenish serpentine or mesh-structure chrysotile or thorn-structure antigorite has the idomorphic outline of olivine enclosed by anthophyllite with a sharp contact. The anthophyllite is also partly altered to talc (Pl. 2, fig. 3). In most cases anthophyllite encloses oval areas now occupied by chrysotile veins with yellowish patches in the middle. The outline of the ovals resembles that of olivine. The serpentine formed after olivine is pale greenish yellow and isotropic (amorphous) and is charged with fine strings of magnetite granules. In some, the core is yellowish green or pale

green and isotropic, while the rims are mesh-structure serpentine. The greenish-yellow portion is, however, not always isotropic and shows mesh structure under crossed nicols. There is thus a transition from isotropic to fibrous serpentine. In mesh-structure serpentine, the median lines from which the chrysotile fibers extend are marked by magnetite granules. There is a regular, anastomosing system; the main veins are thicker, while the veins of the reticulating system are progressively thinner (Pl. 2, fig. 4). In some the median lines lack the filling by magnetite granules and may be isotropic or may show the same polarization color as the fibers. In these characters the rocks resemble those in Sweden described by Du Rietz (1935, p. 246; p. 247, fig. 52). But Du Rietz has observed the chrysotile veins in cracks in fresh olivine grains and is of opinion that the

> "transformation of the olivines, proceeding from the chrysotile veins and attacking the grains partly or altogether is a later phase. In some rock sections this can be seen to be gradual, the olivine at first losing some of its refraction and birefringence. This may happen in the outer part of the grain, sometimes leaving the olivine intact in the centre" (page 247)

He observes that in the Swedish rocks the

> "alteration has gone further, in many cases to a quite amorphous or colloid-like serpentine without any birefringence at all" (page 248) and that "a further stage in the alteration of the amorphous serpentine is a partial brightening of the homogeneous isotropic serpentine which may change the whole rock into aggregates of flame-like serpentine which may be rather like antigorite" (page 249).

In these rocks, however, as will be shown in the description of the secondary minerals, the change is from nonfibrous to fibrous serpentine.

The magnetite strings of the chrysotile fibers extend to the margins of the anthophyllite which are also marked by magnetite granules. The fibers are length slow.

The anthophyllite is pleochroic from a clove brown to green in some sections. Many of these sections show two sets of prismatic cleavages intersecting at angles of 126° and 54° (Pl. 2, fig. 5). When the macrodiagonal of these sections is parallel to the vibration direction of the polarizer, many of the sections show clove brown or tan in one part and green in another part, the two colors grading into each other. Other sections show only green. Sections with one set of cleavage show a pleochroism varying from gamboge yellow to pale tan or lilac. The pleochroism therefore appears to be as follows: X = lilac; Y = clove brown, green; Z = gamboge or canary yellow. The absorption is $b > c = a$. Prismatic sections with one set of cleavages give straight extinction, but in many the extinction is inclined, because the section is not parallel to the *c* axis. There are narrow bands of magnetite granules along the cleavage planes.

Measurement of refractive indices of grains gave the following average results: X = 1.627, Y = 1.628, Z = 1.629.

As the mineral grains are more or less altered, the writer could not obtain a sufficient quantity of pure material for chemical analysis.

Anthophyllite is the most resistant, and in serpentinized rock it is the least-altered. The first stage in its alteration is a change into a colorless mineral with a low-gray polarization color which seems to be a variety of amphibole, with straight extinction in suitable sections. In the next stage, the colorless portion breaks up into needles and bunches of needles of tremolite with yellowish polarization. In many cases the transformation from anthophyllite to tremolite is direct, and the anthophyllite frays out into tremolite needles. The transformation is best seen in sections across the junction between the serpentinized rock and the band of cross fibers (Pl. 3, fig. 1). A common alteration product is serpentine with low-gray polarization color or partly isotropic or fibrous anthophyllite with wavy extinction.

Anthophyllite also alters to chlorite with pale bluish-black polarization color, but the most common alteration is to talc (Pl. 3, fig. 2). In the talc-tremolite schists the talc has formed from anthophyllite, while the tremolite has formed from both the anthophyllite and the pyroxene.

The characteristics of the mineral that suggest a primary origin are: (1) growth of the anthophyllite around primary olivine, in the manner of a reaction rim of orthorhombic pyroxene around olivine, with a lower refractive index, and characteristic pleochroism; (2) the occurrence of oval areas of chrysotile, or a pseudomorph of chrysotile after olivine enclosed

by anthophyllite which is much less altered; (3) alteration of anthophyllite to tremolite and talc; (4) presence of relict anthophyllite in serpentinized peridotites in which all other primary minerals have been altered.

Holland (1895, p. 138) had noted primary anthophyllite in the mica peridotite (mica lamprophyres) of the Darjeeling District and Giridih in India whose description closely resembles that of the mineral of these rocks. Holland noted that, according to Rosenbusch, anthophyllite is common in serpentine rocks. In his *Microscopische Physiographie* (1907, p. 462) Rosenbusch includes anthophyllite peridotite under amphibole peridotites and records an occurrence in Georgia described by King.

Dunham (1950, p. 723) described a nickeliferous norite from New Brunswick in which he found that the orthopyroxenes have been converted into an amphibole which seems to be cummingtonite. It is not strictly a primary mineral, yet it is not a secondary metamorphic mineral.

Anthophyllite with a refractive index γ = ca. 1.61 was synthesized by Bowen and Tuttle (1949, p. 444). They found "that anthophyllite develops as an intermediate phase in the formation of talc from anhydrous materials" and "that pure magnesian anthophyllite has no stable range of existence in the presence of water vapor". Further, they observe "Anthophyllite would form in conditions where the total composition of a mass was such that there was insufficient water to convert the materials to the more hydrous phases talc and serpentine. There would be nothing metastable about this relation" (p. 450). Although in these experiments with a hydrous phase anthophyllite appeared as a passing phenomenon, Bowen and Tuttle envisage the possibility of the formation of anthophyllite in natural rocks (p. 457).

Determination of optic axial angles of anthophyllites from different samples of these rocks gave values of 105°–110° over Z, the majority being 110°, with positive optic sign. One section gave a value of 72° over X with negative optic sign. The birefringence is second-order brown.

The diopsidic pyroxene is colorless with well-developed cleavage lines characteristic of diallage and is in places schillerized. It is invariably characterized by rows of magnetite granules resembling those segregated in serpentinized olivine. In the comparatively less altered rocks, the mineral forms well-developed plates with hypidiomorphic outlines. Figure 3 of Plate 3 shows a prismatic section, more than 2 mm long, which has altered to tremolite needles at one end and to an amphibole at one corner with a sharp boundary against a magnetite plate; the amphibole is pleochroic from green to colorless. That the pyroxene had a sharp contact with olivine is indicated by the sharpness of the junction with oval areas now occupied by fibrous serpentine. Other minerals lying on the margin of the specimen shown in the microphotograph are anthophyllite and chlorite. $Z \wedge c = 36°–39°$. Usually the birefringence is second-order yellow, but second-order blue and indigo are also found; in some, the polarization color is grayish, because the section is nearer the normal to an optic axis.

The 2 V of sections giving bright-yellow

PLATE 1. ASBESTOS QUARRY AND ASBESTOS-BEARING PERIDOTITE

FIGURE 1.—Ultrabasic rock on wall of quarry showing jointing and asbestos veins. The handle of the hammer rests on the veins.

FIGURE 2.—Specimen of ultrabasic rock with bands of cross-fiber asbestos showing branching of asbestos vein and inclusion of nonfibrous rock inside veins (nine tenths natural size).

PLATE 2.—MICROSECTIONS OF DIORITE AND PERIDOTITE SHOWING VARIOUS STAGES IN THE FORMATION OF ASBESTOS VEINS (× 40)

FIGURE 1.—Seive structure of hornblende, transition of hornblende to chlorite and epidote, and rim of leucoxene around ilmenite.

FIGURE 2.—Peridotite with olivine grain in center traversed by cracks and enclosed by anthophyllite on either side.

FIGURE 3.—Altered peridotite showing oval areas filled with mesh-structure serpentine with sharp contact against anthophyllite.

FIGURE 4.—Altered peridotite showing development of veins of chrysotile in serpentinized olivine

FIGURE 5.—Prismatic cleavage in anthophyllite

FIGURE 1

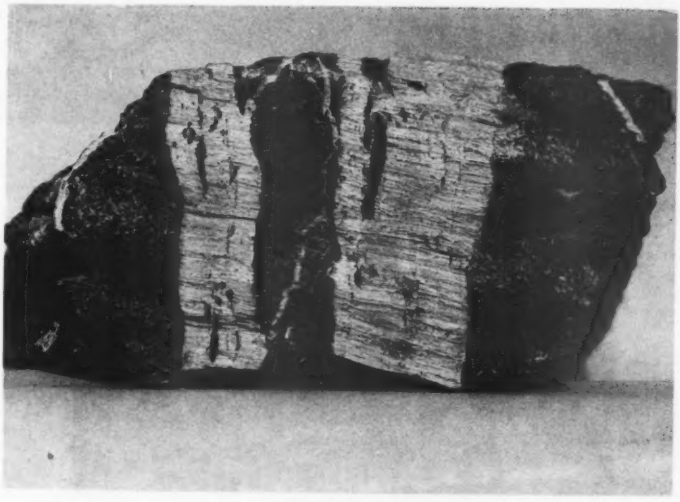


FIGURE 2

ASBESTOS QUARRY AND ASBESTOS-BEARING PERIDOTITE

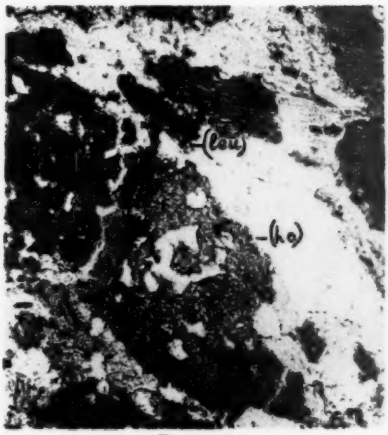


FIGURE 1

FIGURE 2

FIGURE 3

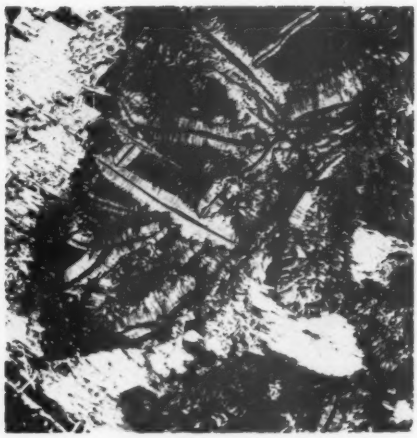


FIGURE 4

FIGURE 5

MICROSECTIONS OF DIORITE AND PERIDOTITE SHOWING VARIOUS STAGES IN THE FORMATION OF ASBESTOS VEINS (40)

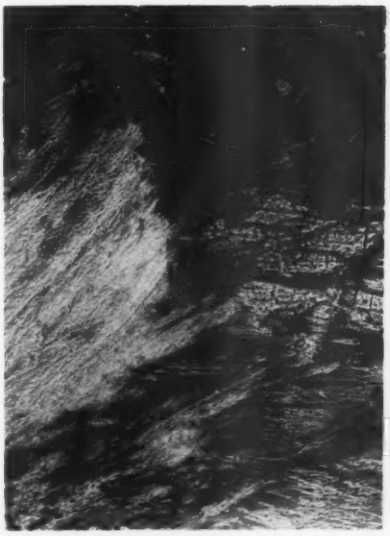


FIGURE 1

FIGURE 2

FIGURE 3

MICROSECTIONS SHOWING FORMATION OF SECONDARY MINERALS (40)

FIGURE 1

FIGURE 2

FIGURE 3

MICROSECTIONS SHOWING THE TRANSFORMATION OF SERPENTINE TO ASBESTOS; REPLACEMENT OF CHRYSOTILE AND EFFECTS OF SHEARING (40)

polarization color of the second order ranges from 55° to 60°; the majority fall near 57.5° with a positive optic sign. A few granules resemble augite in their octagonal outlines, texture, and extinction angle.

Enstatite occurs sparingly in the slides. It has a low polarization color of first order gray, positive optic sign, and an optic axial angle of 55°–60°, the majority being 56°–57°.

The alteration of the pyroxenes is characteristic. Not only are the outlines affected, but the minerals also are altered from within so that torn and shredded skeletal forms are common. The usual alteration is to tremolite and rarely to pleochroic actinolite. The tremolite needles are scattered irregularly through the slide, and some form a felt with ill-defined outlines grading into serpentine. The tremolite needles have a maximum extinction angle of 22° and are length slow. In many cases part of the pyroxene is altered to tremolite needles with yellowish polarization color and to fibrous asbestiform tremolite with grayish polarization.

Magnetite is more abundant than chromite. Primary magnetite forms plates with irregular outlines, rods, and clusters of cubes, while the secondary magnetite forms strings of granules. Chromite grains are reddish brown along their margins, form a brownish-red alteration product, and are gray white in reflected light. The chromite grains are generally found in the anthophyllite or are surrounded by chromium-bearing chlorite.

The two most abundant secondary minerals are serpentine and tremolite. As mentioned, chrysotile develops in the oval, isotropic, serpentinous pseudomorphs after olivine and shows gradual development of mesh structure. In other cases bright streaks appear first in the pale-green isotropic substance, and there are different stages of transition from this to a gratelike structure consisting of minute specks which extinguish alternately at right angles to each other (transverse extinction). This structure was described as "thorn structure" by Bonney and Raisin (1905, p. 702). The thorn structure is similar to the fine felty fibers of antigorite shown by Du Rietz (1935, p. 229, fig. 45). In this when the individual small fibers are slightly larger, they appear as small blades or lamellae with a radial wavy extinction. A third variety of serpentine is a bladed type with wavy extinction resembling antigorite. A fourth type is a feathery variety in which some of the feathers are amorphous. In the fully serpentinized types, areas with mesh structure are interspersed with irregular patches with minute tremolite needles and fibers many of which have a higher polarization color and inclined extinction.

Flett and Hill (1946, p. 64) have described three types of serpentine from the Lizard district—enstatite or bastite serpentine, tremolite serpentine, and dunite serpentine. The tremolite serpentine is dark green when fresh. It

"has always a strong fluxion structure, amounting in places to marked schistosity. It seems to have been injected under strong regional pressures, by which the surrounding schists were being simultaneously folded".

The authors think that

"there is much reason to believe that most of the tremolite in the rocks of the Lizard is paramorphic after pyroxene and olivine".

PLATE 3.—MICROSECTIONS SHOWING FORMATION OF SECONDARY MINERALS (× 40)

FIGURE 1.—Junction between nonfibrous serpentinized peridotite and asbestos veins. Note the passage of anthophyllite to tremolite.

FIGURE 2.—Formation of talc from anthophyllite

FIGURE 3.—Diallage plate showing fine cleavage; diallage changing into amphibole at the right-hand corner against magnetite.

PLATE 4.—MICROSECTIONS SHOWING THE TRANSFORMATION OF SERPENTINE TO ASBESTOS; REPLACEMENT OF CHRYSOTILE, AND EFFECTS OF SHEARING (× 40)

FIGURE 1.—Thorn-structure serpentine swamped by tremolite needles (t); a piece of unaltered relic anthophyllite (a) near top.

FIGURE 2.—Showing replacement of chrysotile by two patches of tremolite in an asbestos band

FIGURE 3.—Sheared serpentinized rock showing flow structure of asbestos fibers and magnetite granules. Note relic anthophyllite (a).

The tremolite serpentine had attained its present structure and mineral composition when the injection and folding movements concluded.

The chlorite is pale green, and some of it is strongly pleochroic from pale green to light pink or pinkish violet. Some of it is isotropic, but it generally gives a washed, inky, polarization color. In the serpentinized rocks the chlorite has a plumose or spherulitic extinction or has a feathery appearance with undulatory extinction and extends out of the anthophyllite. The pleochroic chlorite has an optic axial angle of 0° and a positive sign. Its characteristics agree with those of the chromium-bearing chlorite kammererite. The granules and patches of chromite are generally set in a matrix of this chlorite. While the chromium in the chlorite might have been derived from the chromite, the chlorite seems rather to have formed after anthophyllite; a transitional stage is marked by a brownish product, with separation of strings of iron ores, which is pleochroic from brown to dark, and hence the chromium must have been in the anthophyllite.

The serpentinized peridotites throw light on the manner of transformation of the whole rock into "tremolite asbestos" and of the non-fibrous serpentine to fibrous serpentine. The change from unaltered rock to a talc schist through serpentinized varieties is accompanied by a decrease of specific gravity from 2.86 to 2.55 in the talc schist.

The partly altered peridotite shows tremolite, anthophyllite, antigorite, chlorite, and talc. Antigorite is flaky with wavy and wandering extinction. The chlorite has also a wavy plumose extinction and may with difficulty be distinguished from antigorite; but in ordinary light the latter is colorless, and the former is slightly green. The chlorite is also faintly pleochroic from light golden yellow to pale green or bluish green. It also includes in places small cubes of magnetite or chrome magnetite. Part of the chloritic patches is isotropic, and a patch of elongated fibrous isotropic material is derived from anthophyllite. The original positions of olivine are recognized by "thorn structure" serpentine. The tremolite needles and fibers are recognized by their higher relief and birefringence.

A common but significant feature of these rocks is that in ordinary light clear, light-greenish patches are surrounded by a felty mass of tremolite needles. The light-greenish patches show thorn structure under crossed nicols. The boundary between them and the felty mass of tremolite is irregular, and the tremolite needles replace the felty mass (Pl. 4, fig. 1). The manner of transformation of chrysotile or antigorite to tremolite is clearer in the sections taken from the band of the cross fibers. Though the cross fibers seem to be chrysotile, chemical analysis shows a peculiar composition which is due to incomplete tremolitization of the chrysotile fibers. Figure 2 of Plate 4 shows two patches of tremolite, distinguished by their long fibers, with inclined extinction and higher interference color of yellow, spreading over a thorn-structure antigorite by progressive replacement of the latter.

Sections across the junction of the cross fibers and the non-fibrous serpentine (Pl. 3, fig. 1) show the transformation of anthophyllite into tremolite fibers and of serpentinized olivine into chrysotile fibers. Other sections between fibrous tremolite and serpentine show that the junction is irregular and that the fibers extend from the relics of anthophyllite thus suggesting the transformation of the one into the other. That this is later than the serpentinization is proved by the incipient fiberization in the greenish serpentine, and by the coalescence of the chrysotile fibers in cross fiber veins with fine strings of magnetite granules between the fibers.

Microscopic examination of the sheared and bent varieties of the serpentinized rocks shows the effect of stress in the parallel orientation it has imparted to all the constituents (Pl. 4, fig. 3). The tremolite fibers show the influence of stress in their flowlike texture. Under crossed nicols elongated patches of isotropic chlorite are found mixed with tremolite fibers. The released granules of magnetite form strings and clots. Other minerals are relics of anthophyllite, greenish hornblende, and actinolite. The microphotograph (Pl. 4, fig. 3) shows a relict anthophyllite with well-marked cleavages at an angle to the flow lines of the fibers. Other such relics have random orientation some of which have their longer direction at right angles to the direction of elongation. This shows that the

stress operated toward the closing stages of serpentinization and continued during fiberization.

A few chalcopyrite pieces were found enveloped in tremolite fibers which formed nodules in the massive fibrous tremolite.

Chemical Composition of Younger Ultrabasic and of Tremolite Asbestos

The chemical composition of the typical peridotite which has been serpentinized least is given (Table 1) for comparison with the anlyses of a hornblende-augite olivinite quoted by Johannsen (1949, p. 417) and of a serpentinized peridotite from the Merrimac area described by Hietanen (1951, p. 577). (1) is an analysis of "M/185", a partially serpentinized peridotite from Manpur, Dhalbhum, Singhbhum District, Bihar. Analyst—W. H. Herdsman, Glasgow. The norm consists of hypersthene 40.24 per cent, diopside 22.5 per cent, and olivine 11.7 per cent. The hypersthene of the norm probably represents most of the anthophyllite of the mode.

The mode of (2) is olivine 73 per cent, pyroxene 14 per cent, amphibole 10 per cent, and pyrrhotite 3 per cent. The norm of (3) is pyroxene 49.6 per cent, and olivine 26.5 per cent. The petrographical description of (3) agrees closely with that of the analyzed specimen from Manpur. It consists of large tremolite plates and remnants of olivine containing small tremolite needles embedded in a mass of antigorite. The rock is associated with serpentine and talc schists which represent more nearly complete alteration products of the same intrusive.

The chemical composition of the tremolite asbestos is

SiO_2	57.36
FeO	5.54
CaO	12.32
MgO	21.94
H_2O^- 105°C	0.18
H_2O^+ 105°C	1.22
	98.56

Analyst W. H. Herdsman, Glasgow.

Origin of Asbestos

The problem of the origin of the asbestos presents the question of (1) the origin of serpentinization, (2) transformation of the non-fibrous serpentine into fibrous asbestos, and (3) mass transformation of chrysotile into tremolite. Opinion is sharply divided about the source of serpentinizing solutions. Benson (1918, p. 693) believed that the solutions represented residual liquids of the ultrabasic magma; others suggested younger acid intrusives. Hess (1933, p. 652) was the main proponent of the former hypothesis in its modern form and considered serpentinization an "autometamorphic" reaction. According to him the residual liquid was not merely a highly aqueous (hydrothermal) solution but a highly siliceous solution containing relatively less water for which he proposed the term "hypo-hydrous". The work of Haapala (1936, p. 71) led to a similar conclusion. Du Rietz (1935, p. 233) did not observe any "facts supporting the derivation of the serpentinizing solutions from peridotite magmas". He noted that "among the peridotites in mica schists and gneisses serpentinized ultrabasics are local occurrences, and are often directly connected with granite and pegmatite intrusions". The serpentinization of the peridotites in phyllites, according to him, might be due to waters emitted from acid magmas intruded in an earlier phase.

The theory proposed by Hess fails to explain the occurrence of "kernels of nearly isotropic serpentine in an interlacing network of veinlets

Table 1.—Chemical Analyses

	(1)	(2)	(3)
SiO_2	43.84	40.35	46.52
Al_2O_3	3.46	3.75	1.81
FeO	5.94	9.86	3.52
Fe_2O_3	7.88	3.53	5.39
TiO_2	0.34	2.12	0.13
MnO	0.24	0.20	0.11
CaO	5.92	4.46	8.26
MgO	25.33	25.69	27.52
K_2O	0.68	0.80	0.08
Na_2O	0.12	3.14	1.33
H_2O −105°C	0.70	0.83	0.14
H_2O +105°C	5.79	5.28	4.82
CO_2	nil	trace	0.71
P_2O_5	trace	0.25	0.00
BaO, SrO	n.d.	0.06	.....
Total	100.24	99.32	100.34

of fibrous chrysotile" (Turner, 1948, p. 131). The Hess theory requires the "attack of the olivine grains by a residual liquid present in the interstices between the grains", a process which must stop after the network of chrysotile fibers encloses kernels of unaltered olivine.

The results of the experimental work of Bowen and Tuttle (1949) have settled the problem, and Hess has accepted them with some reservations (Hess, 1949, p. 857). As a result of their experimental investigations, Bowen and Tuttle concluded that serpentinization is not an autometamorphic process but is "the work of solutions of extraneous origin which have probably reheated the peridotite". Even if the serpentinization belongs to the same cooling cycle, the aqueous medium is of extraneous origin.

In the present case both field facts and microscopic evidence show that (1) serpentinization is a post-magmatic phenomenon, and (2) the solutions responsible for the changes are not residual in the autometamorphic sense. Serpentinization has not only proceeded from cracks which are represented by veins of fibers, but has affected the olivine grains uniformly even in the compact rock. The distribution of serpentinization is irregular and is related neither to the margin of the ultrabasic bodies nor to their cores. Its relation with the joint planes in the ultrabasic is further evidence of its post-consolidation age.

On the other hand there was no injection of younger granite or pegmatite veins either in the ultrabasic or in the neighborhood. The adjoining gneisses are older country rocks and could not have supplied the solutions. In fact, in the contact rock examined, there is considerable tremolite, probably formed by the migration of materials from the ultrabasic at the time of its serpentinization. What then is the source of the serpentinizing solutions? The source must be in some deeper zone below the ultrabasic intrusions possibly connected with the same magmatic cycle.

The shearing movements which affected the ultrabasic rocks at this stage offered channels for the ascent of the liquids. These movements must have continued even after the formation of the fibers for some of the serpentinized peridotites have been sheared (Pl. 4, fig. 3). These shearing movements may represent the last phase of the more extensive movements which caused the zone of thrusts in eastern Singhbhum.

This particular occurrence also throws light on the manner of vein formation. Although some of the veins of tremolite occupy joint planes, as if deposited from a solution, the majority form cross-fiber and slip-fiber veins. The irregular thickness of the veins and their branching into progressively thinner veinlets in all directions show that the veins do not represent pre-existing fissures. Dresser (1913, p. 65) objected to the fissure-filling hypothesis on similar grounds and suggested replacement by solutions which worked outward from the tight cracks which now form the central fissure of the vein. Taber objected to the replacement hypothesis of Dresser; he suggested that the vein-forming fluids penetrated through pores and microscopic openings in the rock, and cross-fiber veins formed by a process of lateral secretion. He attributed the fibrous structure to physical conditions which limited crystal growth to a single direction (Taber, 1918, p. 74). Cooke (1936, p. 355), further expanding the ideas of Taber, postulated the movement of the asbestos-forming fluids through pores and fissures in the rock, and growth of the fibers at both ends by the supply of material through the pores. The main feature of Cooke's hypothesis is the constancy of the ratio of serpentine zones flanking the veins to the width of the veins.

In the rocks under study microscopic and field evidences favor the replacement hypothesis. The irregular boundary of the vein and the interpenetration of the fibers and the serpentinized wall rock as shown in Figure 2 of Plate 1 show that under the influence of stress the nonfibrous serpentine has been transformed into fibrous serpentine. The similarity in composition between serpentine and chrysotile has been advanced as an argument against the transformation of the one into the other (Taber, 1916, p. 73), but, as Bateman (1918, p. 680) has pointed out, serpentines of diverse physical and optical properties form in response to differing physical and chemical conditions.

Proud and Osborne (1952, p. 13) have shown the influence of stress environment in the genesis of chrysotile with reference to a part of

the Great Serpentine Belt of New South Wales. According to these authors, fiber development took place in two stages: (1) the development of rotational stress caused a distinctive fracture pattern and led to a second intrusion of dunite-harzburgite, which was serpentinized by solutions of extraneous origin; (2) the final shearing process caused injection of dolerites and gabbros.

The chemical reactions which led to the association of tremolite and calcite and the transformation of chrysotile into tremolite might have been:

(1) $\underset{\text{Diopside}}{5CaMgSi_2O_6} + H_2O + 3CO_2$

$= \underset{\text{Tremolite}}{Ca_2Mg_5Si_8O_{22}(OH)_2} + 3CaCO_3 + 2SiO_2$

or,

$5CaMgSi_2O_6 + H_2O$

$= Ca_2Mg_5Si_8O_{22}(OH)_2 + 3CaO + 2SiO_2$

Both the above reactions seem to have taken place since calcite is of very limited occurrence. The second reaction released lime which caused lime metasomatism by which chrysotile was transformed into tremolite. This was aided by the silica liberated in the process and probably supplemented by the influx of more silica from deeper sources.

(2) $2\{\underset{\text{Chrysotile}}{Mg_3(OH)_4Si_2O_5} + 2CaO + 4SiO_2$

$= \underset{\text{Tremolite}}{Ca_2Mg_5Si_8(OH)_2O_{22}} + 3H_2O + MgO$

According to Turner (1933, p. 257), the lime-silicate rocks in the peridotite belt of South Westland originated by the "alteration under pressure of basic dike rocks, by aqueous solutions rich in lime" by the same waters which caused the serpentinization, and the lime content was derived from the monoclinic pyroxenes during serpentinization.

Age of Intrusive, its Tectonic Setting, and Relation to Other Ultrabasic Occurrences in Singhbhum

Ultrabasic rocks occur as small isolated masses widely scattered in eastern and southern Singhbhum. According to Dunn (1929, p. 96) they are related and belong to one period of intrusion. The largest bodies, described by Fermor (1919), occur as thick laccolithic intrusions in the Iron Ore Series (Dharwars) near Jojohatu and have been affected by later stages of folding of the Iron Ore Series. These consist of saxonites, dunite, pyroxenite, and a specimen of lehrzolite. They have been altered largely to serpentine carrying chromite and to talc schist and schistose serpentine near the margins where shearing is greatest. Since the granite is intrusive into the Iron Ore Series, the ultrabasics were considered to be pre-granite but later than the metamorphism and severe shear movements of the Iron Ore Series. They have suffered shearing to the same degree as the granite. According to Fermor, serpentinization has been effected by silicious solutions.

The ultrabasic bodies in the Singhbhum granite and other granite gneisses are considered inclusions. To this class, according to Dunn, belong the ultrabasic bodies found at Bara Bana (22°37′:85°56′) and Rajnagar in the Singhbhum district, and the Manpur occurrence, all of which carry asbestos.

Jones (1934, p. 178) considered the ultrabasic rocks of South Singhbhum and the adjoining parts of Orissa intrusive into the granites.

In later work on East Singhbhum Dunn and Dey considered the ultrabasic dike rocks between Ghatsila (22°34′:86°29′) and Mushaboni (22°31′:86°28′) part of a younger suite because they are unmetamorphosed (1942, p. 361, 426). They concluded that there are ultrabasic intrusions of more than one age. The writer concludes that the Manpur ultrabasic intrusions are younger than the granite because of: (1) structural relation with the enclosing gneiss, and the downward expansion in one of the quarries, (2) contact effect on the gneiss in one of the quarries, and (3) comparative freedom of the ultrabasic from regional metamorphism which has affected the gneisses.

The ultrabasics are less sheared than the gneisses and granites. They might owe their intrusion to the same earth movements which have given rise to the zone of thrusts and shears. The changes which caused serpentinization and formation of asbestos mark the closing stages of these movements.

The above criteria also indicate that the intrusion was a solid mass which had been forced

up by earth movements. The location of ultrabasic and serpentine intrusions in zones of crustal deformation has been noted in the West Indies (Hess, 1938, p. 332–333), the island arcs of North Pacific (Hess, 1948), and Australia. In Cuba Thayer and Guild (1947) have shown the intrusive association in solid state of the ultrabasic with overthrusts.

Sen Gupta (1942) studied the asbestos deposits of Rajnagar in Seraikela, about 15 miles west of Manpur; these originated from an ultrabasic which belongs to the same suite. The asbestos is tremolite, but Sen Gupta believed that it was chrysotile because of mesh structure in serpentine pseudomorphs after olivine. He explained the unusual length of the fibers as due to coalescence of seams of cross-fiber veins. The present investigation suggests that the asbestos is mainly tremolitic in Rajnagar area also.

Conclusion

The writer believes the ultrabasic bodies found inside the diorite and granites of the Manpur area are intrusive; that serpentinization was a post-consolidation phenomenon, and that the source of the serpentinizing solutions lay in deep parts of the crust. Serpentinization was caused by replacement, and the nonfibrous serpentine was transformed into fibrous serpentine under the influence of stress.

References Cited

Amin, M. S., 1952, Metamorphic differentiation of talc-magnesite-chlorite rocks in Shetland: Geol. Mag., v. 89, no. 2, p. 97–106.

Bateman, Alan M., 1918, An Arizona asbestos deposit: Econ. Geology, v. 23, p. 663–680.

Benson, W. N., 1918, The origin of serpentine, a historical and comparative study: Am. Jour. Sci. 4th ser, v. xlvi, p. 693–730.

Bonney, T. G., and Raisin, C. A., 1905, The microscopic structure of minerals forming serpentine, and their relation to its history: Geol. Soc. London Quart. Jour., v. 61, p. 694–714.

Bowen, N. L., and Tuttle, O. F., 1949, The system MgO-SiO_2-H_2O: Geol. Soc. America Bull., v. 60, p. 439–460.

Cooke, H. C., 1936, Asbestos deposits of Thetford district, Quebec: Econ. Geology, v. 31, p. 355.

Davis, G. L., and Hess, H. H., 1949, Radium content of ultramafic igneous rocks, Part II. Geological and chemical implications: Am. Jour. Sci., v. 247, p. 856–882.

Dresser, J. A., 1913, Preliminary report on the serpentine and associated rocks of southern Quebec: Geol. Survey Canada Mem. 22, 103 pages.

Dunham, K. C., 1950, Nickeliferous norite of New Brunswick: Am. Mineralogist, v. 35, p. 723.

Dunn, J. A., 1929, The geology of north Singhbhum including parts of Ranchi and Manbhum districts: Geol. Survey India Mem., v. liv, 166 pages.

Dunn, J. A., and Dey, A. K., 1942, The geology and petrology of eastern Singhbhum and surrounding areas: Geol. Survey India Mem., vol. LXIX, pt. 2, p. 281–456.

Du Rietz, T., 1935, Peridotites, serpentines and soapstones of North Sweden: Geol. Foren. Stockholm, Band 57, p. 135–260.

Fermor, L. L., 1919, General Report, Geological Survey of India: Geol. Survey India Rec., v. L, p. 10–12.

Flett, J. S., and Hill, J. B., 1946, Geology of the Lizard and Meneage: Geol. Survey Great Britain, England and Wales Mem., 2d. ed., by Sir John Smith Flett, 208 pages.

Haapla, P., 1936, On the serpentine rocks in northern Karelia: Comm. Geol. Finlande Bull., no. 114, p. 5–80.

Hess, H. H., 1933, The problem of serpentinization and the origin of certain chrysotile asbestos talc and soapstone deposits: Econ. Geology, v. 28, p. 634–657.

——1948, Major structural features of the Western North Pacific: Geol. Soc. America Bull., v. 59, p. 417–446.

Hietanen, A., 1951, Metamorphic and igneous rocks of the Merrimac Area, Plumas National Forest, California: Geol. Soc. America Bull., v. 62, p. 565–608.

Holland, T. H., 1895, On highly phosphatic mica peridotites intrusive in the Lower Gondwana rocks of Bengal: Geol. Survey India Rec., v. XXVII, pt. 4, p. 138.

Johannsen, A., 1949, A descriptive petrography of the igneous rocks: Chicago, Univ. Chicago Press, v. 4, pt. 2, 511 pages.

Jones, H. C., 1934, The iron ore deposits of Bihar and Orissa: Geol. Survey India Mem., v. 63, pt. II, p. 167–302.

Proud, J. S., and Osborne, G. D., 1952, Stress environment in the genesis of chrysotile with special reference to the occurrence at Woodsreef near Barraba, New South Wales: Econ. Geology, v. 47, p. 13–23.

Read, H. H., 1934, The metamorphic geology of Unst in the Shetland Islands: Geol. Soc. London Quart. Jour., v. xc, p. 637–688.

Rosenbusch, H., 1907, Microscopische physiographie der mineralien und gesteine, E. S. Verlagshandlung, Stuttgart, 716 pages.

Sen Gupta, K. K., 1942, The asbestos deposits of Rajnagar, Seraikela State: Geol. Min. Met. Soc. India Quart. Jour., v. xiv, p. 91–98.

Taber, S., 1916, The genesis of asbestos and asbestiform minerals: Am. Inst. Min. Engineers Trans., v. lvii, p. 62–69.

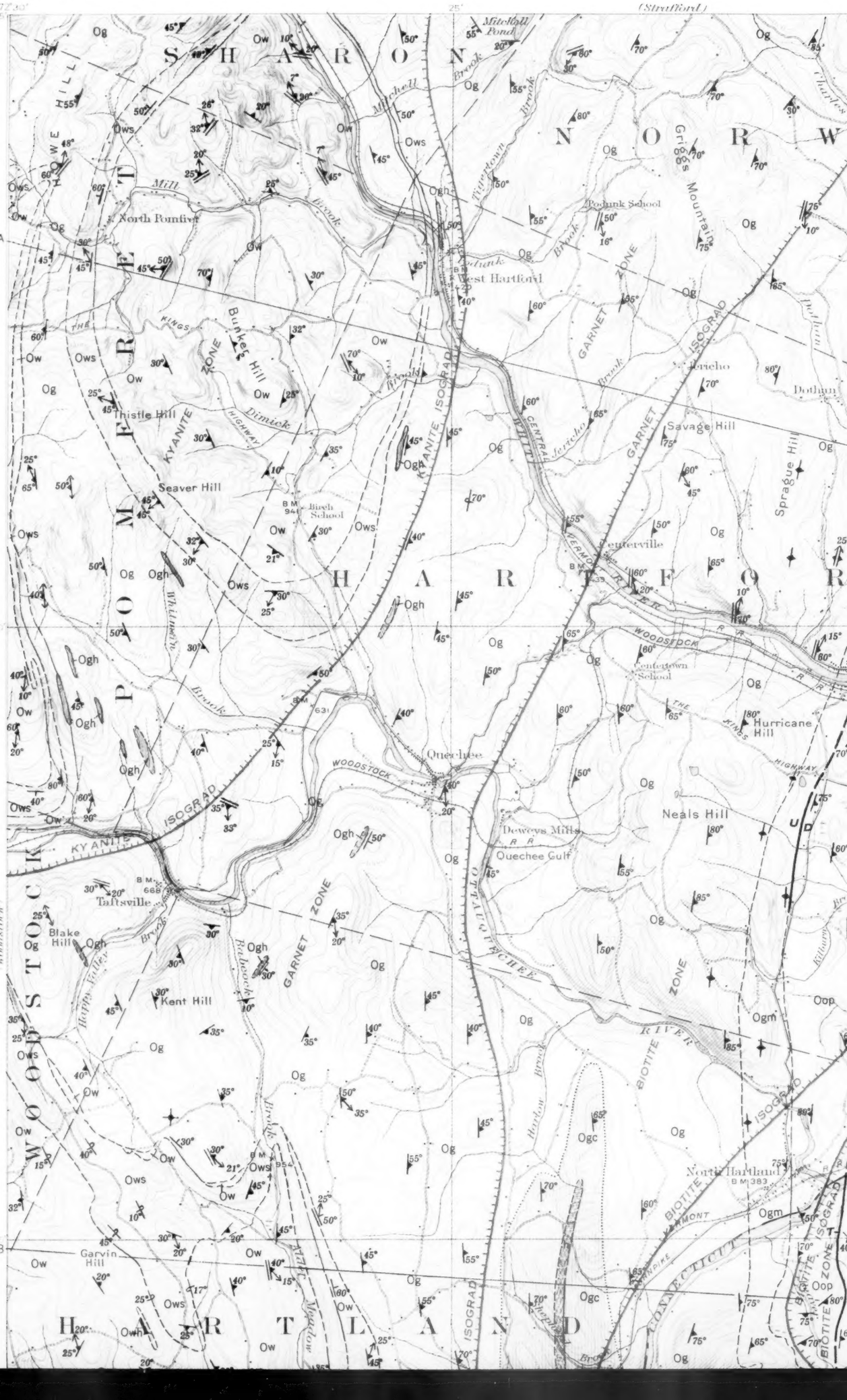
Thayer, T. P., and Guild, P. W., 1947, Thrust faults and related structures in eastern Cuba: Am. Geophys. Union Trans., v. 28, p. 919–930.

Turner, F. J., 1948, Mineralogical and structural evolution of metamorphic rocks: Geol. Soc. America Mem. 30, 342 pages.

Patna University, Patna 5, Bihar, India

Manuscript Received by the Secretary of the Society, August 17, 1953

NEW HAMPSHIRE STATE PLANNING
AND DEVELOPMENT COMMISSION



VERMONT-NEW HAMPSHIRE
HANOVER QUADRANGLE

LEGEND

IGNEOUS AND METAMORPHIC ROCKS

DEVONIAN

lg / lb — **Lebanon granite**

(lg, Lebanon granite, a medium- to coarse-grained pink, subporphyritic, somewhat granulated granite and gneissic granite, composed of microcline, quartz, oligoclase, biotite, epidote, and muscovite. lb, border gneiss, quartz-diorite, composed of oligoclase, quartz, biotite, epidote, microcline, and muscovite.)

a — **Amphibolite**

(Intrusive amphibolite: dark-green well-lineated amphibolite composed of hornblende, andesine, and epidote.)

Dl — **Littleton formation**

(Black to gray quartz-mica schist, mica-quartz schist, and micaceous quartzite; biotite porphyroblasts locally common.)

Metamorphosed diabase

(Numerous sills and dikes of medium-grained dark-green meta-diabase composed of sodic plagioclase, hornblende or chlorite, and epidote. Within the chlorite zone these intrusives are greenschists; within the garnet zone they are amphibolites.)

gn — **Gneiss**

(At White River Junction a lit-par-lit gneiss injected into the Post Pond volcanic member of the Orfordville formation; banded gray gneiss composed of quartz, albite, and minor muscovite, chlorite, and epidote. Northeast of Plainfield a well-lineated gneiss composed of quartz, albite, chlorite, and epidote.)

h — **Hornblende gneiss**

(Coarse-grained intrusive hornblende gneiss composed of hornblende, oligoclase, and biotite. A 200-foot thick tourmalinized sill injected into the Gile Mountain formation.)

ORDOVICIAN (?)

Oop / Ooc / Oo / Oov / Ooh / Oov / Oo / Oog — **Orfordville formation**

(West of the Ammonoosuc thrust the Orfordville formation consists chiefly of chlorite schist of the Post Pond volcanic member, Oop, and of black phyllite, Oo. East of the Ammonoosuc thrust the Orfordville formation consists chiefly of hornblende schist, amphibolite, and feldspathic schist of the Post Pond volcanic member, Oop, and of gray to black mica schist and quartz-mica schist, Oo, in which garnet porphyroblasts are common. The Hardy Hill quartzite, Ooh, is a thin quartzite and quartz meta-conglomerate in the lower part of the formation. Small lentils of quartz-calcite schist and marble, Ooc, and of hornblende or chlorite schist, Oov, are locally present, in the southeastern part of the map area. Orfordville schists peripheral to the Lebanon granite have been metasomatized to feldspathic gneiss, Oog.)

Ogm / Og / Ogc / Ogh / Og — **Gile Mountain formation**

(West of the garnet isograd the rocks are tan to gray quartz-mica schist and micaceous quartzite, generally containing garnet and biotite porphyroblasts, Og. East of the garnet isograd the same rocks, Og, are finer-grained, and lack garnet. The easternmost member of the formation, the Meetinghouse slate member, Ogm, is a black, thin-bedded phyllite and mica schist. Black calcite-quartz schist, Ogc, is widespread in the south-central part of the map area. Locally there are lentils of dark-green hornblende schist, Ogh, of probable volcanic origin.)

Ows

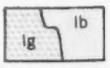
NEW HAMPSHIRE STATE PLANNING
AND DEVELOPMENT COMMISSION
(Randolph)
72°30'
43°45'
25'
(Strafford)
40'
35'
A
B
(Woodstock)
SHARON
NORWICH
POMFRET
HARTFORD
HARTLAND
WOODSTOCK
North Pomfret
West Hartford
Quechee
Taftsville
North Hartland
Seaver Hill
Thistle Hill
Burkett Hill
Savage Hill
Hurricane Hill
Neals Hill
Blake Hill
Kent Hill
Garvin Hill
Sprague Hill
Centerville
Dothan
Jericho
Dewey's Mills
Quechee Gulf
Mill Brook
Happy Valley Brook
Barrows Brook
Hartland Four Corners
Kings Highway
Dimick Highway
Woodstock
Centerville School
Birch School
Podunk School
Mitchell Pond
Mitchell Brook
Middle Brook
Jericho Brook
Hartland Four Corners
Gris Mountain
KYANITE ISOGRAD
GARNET ISOGRAD
STAUROLITE-KYANITE ISOGRAD
STAUROLITE ISOGRAD
GARNET ZONE
BIOTITE ISOGRAD
BIOTITE ZONE
Ogm
Ogc
Oop
Ogh
Ow
Ows
Og

VERMONT-NEW HAMPSHIRE
HANOVER QUADRANGLE

LEGEND

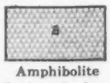
IGNEOUS AND METAMORPHIC ROCKS

DEVONIAN



Lebanon granite

(lg, *Lebanon granite, a medium- to coarse-grained pink, subporphyritic, somewhat granulated granite and gneissic granite, composed of microcline, quartz, oligoclase, biotite, epidote, and muscovite.* lb, *border gneiss; quartz diorite, composed of oligoclase, quartz, biotite, epidote, microcline, and muscovite.*)

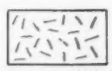


Amphibolite

(*Intrusive amphibolite; dark-green well-lineated amphibolite composed of hornblende, andesine, and epidote.*)

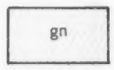
Littleton formation

(*Black to gray quartz-mica schist, mica-quartz schist, and micaceous quartzite; biotite porphyroblasts locally common.*)



Metamorphosed diabase

(*Numerous sills and dikes of medium-grained dark-green meta-diabase composed of sodic plagioclase, hornblende or chlorite, and epidote. Within the chlorite zone these intrusives are greenschists; within the garnet zone they are amphibolites.*)



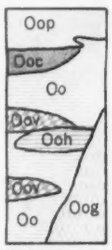
Gneiss

(*At White River Junction a lit-par-lit gneiss injected into the Post Pond volcanic member of the Orfordville formation; banded gray gneiss composed of quartz, albite, and minor muscovite, chlorite, and epidote. Northeast of Plainfield a well-lineated gneiss composed of quartz, albite, chlorite, and epidote.*)

Hornblende gneiss

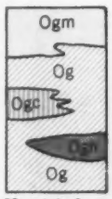
(*Coarse-grained intrusive hornblende gneiss composed of hornblende, oligoclase, and biotite. A 200-foot thick tourmalinized sill injected into the Gile Mountain formation.*)

ORDOVICIAN (?)



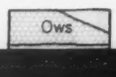
Orfordville formation

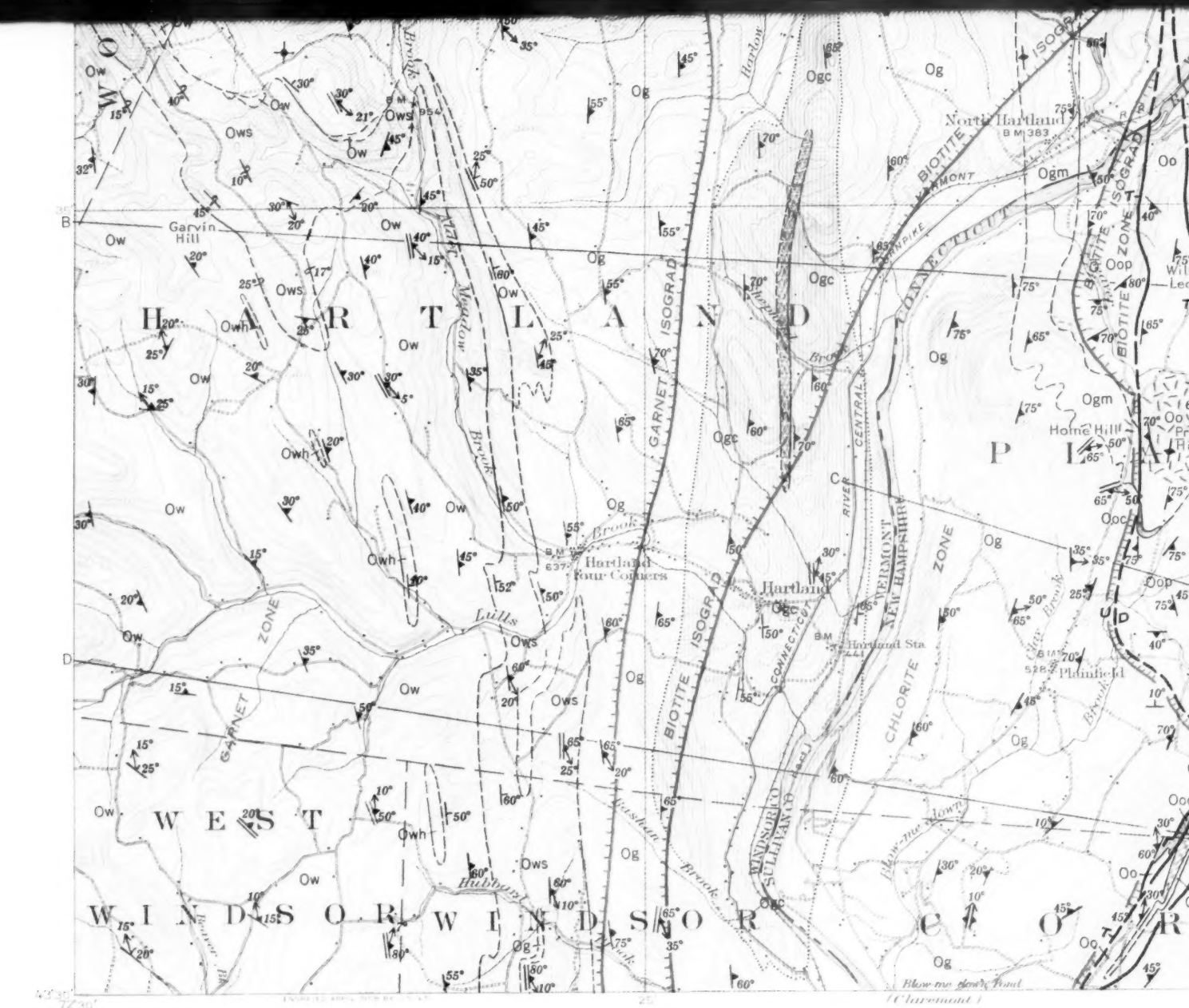
(*West of the Ammonoosuc thrust the Orfordville formation consists chiefly of chlorite schist of the Post Pond volcanic member,* Oop, *and of black phyllite,* Oo. *East of the Ammonoosuc thrust the Orfordville formation consists chiefly of hornblende schist, amphibolite, and feldspathic schist of the Post Pond volcanic member,* Oop, *and of gray to black mica schist and quartz-mica schist,* Oo, *in which garnet porphyroblasts are common. The Hardy Hill quartzite,* Ooh, *is a thin quartzite and quartz meta-conglomerate in the lower part of the formation. Small lentils of quartz-calcite schist and marble,* Ooc, *and of hornblende or chlorite schist,* Oov, *are locally present, in the southeastern part of the map area. Orfordville schists peripheral to the Lebanon granite have been metasomatized to feldspathic gneiss,* Oog.)

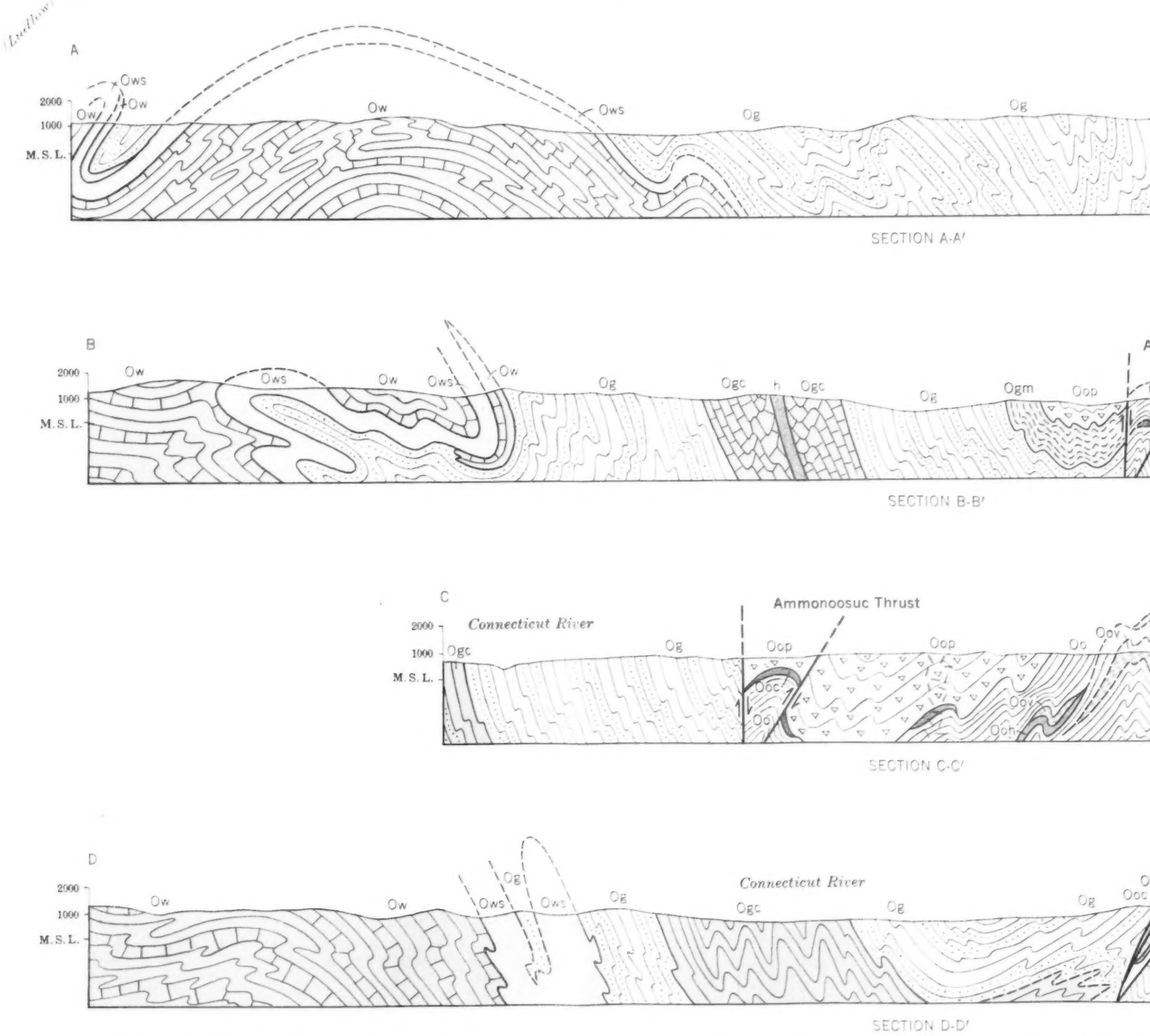


Gile Mountain formation

(*West of the garnet isograd the rocks are tan to gray quartz-mica schist and micaceous quartzite, generally containing garnet and biotite porphyroblasts,* Og. *East of the garnet isograd the same rocks,* Og, *are finer-grained, and lack garnet. The easternmost member of the formation, the Meetinghouse slate member,* Ogm, *is a black, thin-bedded phyllite and mica schist. Black calcite-quartz schist,* Ogc, *is widespread in the south-central part of the map area. Locally there are lentils of dark-green hornblende schist,* Ogn, *of probable volcanic origin.*)







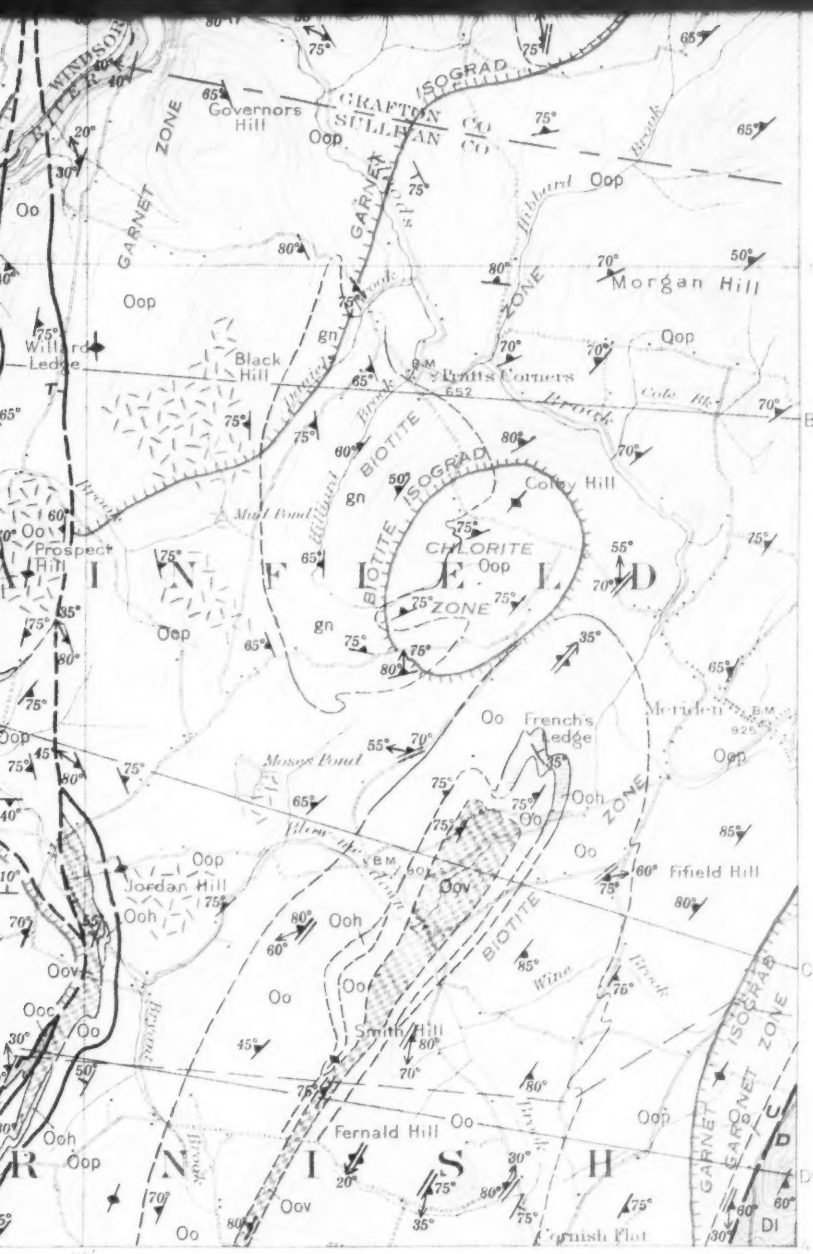
GEOLOGIC MAP AND STRUCTURE SECTIONS OF THE HANOVER QUADRAN

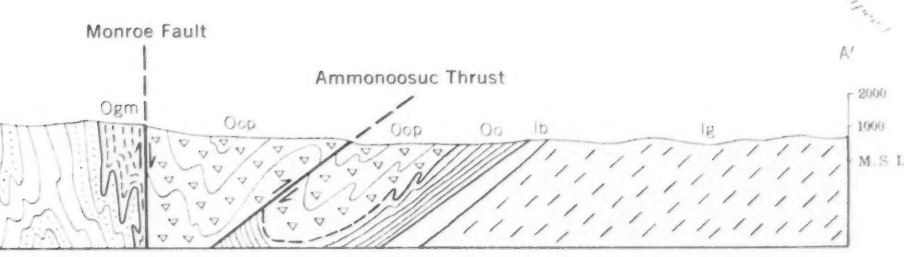
Topographic base by U. S. Geological Survey. Surveyed in 1905 and 1906

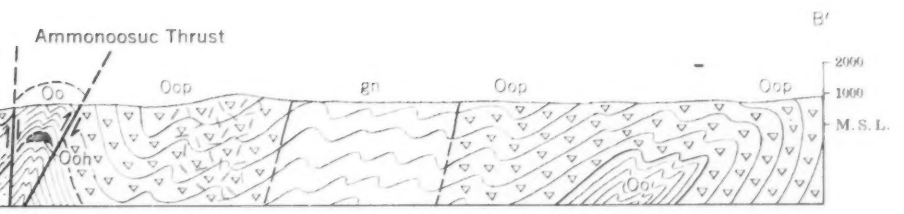
Scale 1/62500

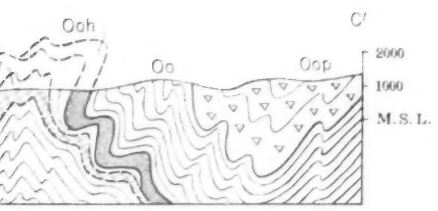
Contour interval 20 feet.

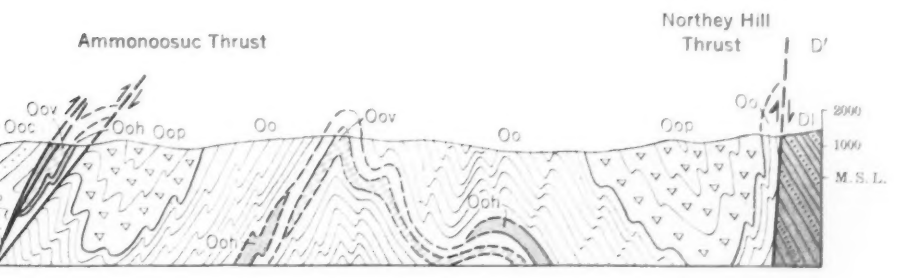
Datum is mean sea level

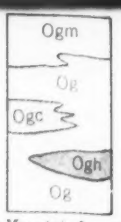






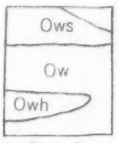






Gile Mountain formation

(West of the garnet isograd the rocks are tan to gray quartz-mica schist and micaceous quartzite, generally containing garnet and biotite porphyroblasts, Og. East of the garnet isograd the same rocks, Og, are finer-grained, and lack garnet. The easternmost member of the formation, the Meetinghouse slate member, Ogc, is a black, thin-bedded phyllite and mica schist. Black calcite quartz schist, Ogc, is widespread in the south-central part of the map area. Locally there are lentils of dark-green hornblende schist, Ogg, of probable volcanic origin.)



Waits River formation

(Main part of the formation, Ow, consists of brown-weathering calcite-quartz schist, with lesser amounts of mica schist and mica quartz schist containing garnet porphyroblasts. Standing Pond amphibolite member, Ows, is chiefly a dark-green, garnetiferous, hornblende gneiss. Lentils of hornblende schist, Owh, are present locally.)

NOTE. Albee and Ammonoosuc formations (Ordovician?) and Clough and Fitch formations (Silurian) are cut out by the Northey Hill thrust where the Orfordville and Littleton formations are in juxtaposition. The Meetinghouse slate is, in part, the probable equivalent of the lower portion of the Orfordville formation.

METAMORPHIC ZONES

Metamorphic zones shown by isograds, hachured on the high-intensity side. Rocks east of the Ammonoosuc thrust in the southern portion of the area are in the biotite zone; those west of the thrust are in the chlorite zone.

CONTACTS

Accurate formation contact

Indefinite formation contact

Concealed or inferred contact

STRUCTURE SYMBOLS

25° — Strike and dip of bedding

50° — Strike and dip of overturned beds

30° — Strike and dip of foliation or schistosity

Strike of vertical foliation or schistosity

Horizontal foliation or schistosity

50° →40° — Foliation or schistosity, with strike and plunge of lineation

20° ←30° — Strike and dip of axial plane of minor fold, with strike and plunge of fold axis

T — Thrust fault; (T, upper plate)

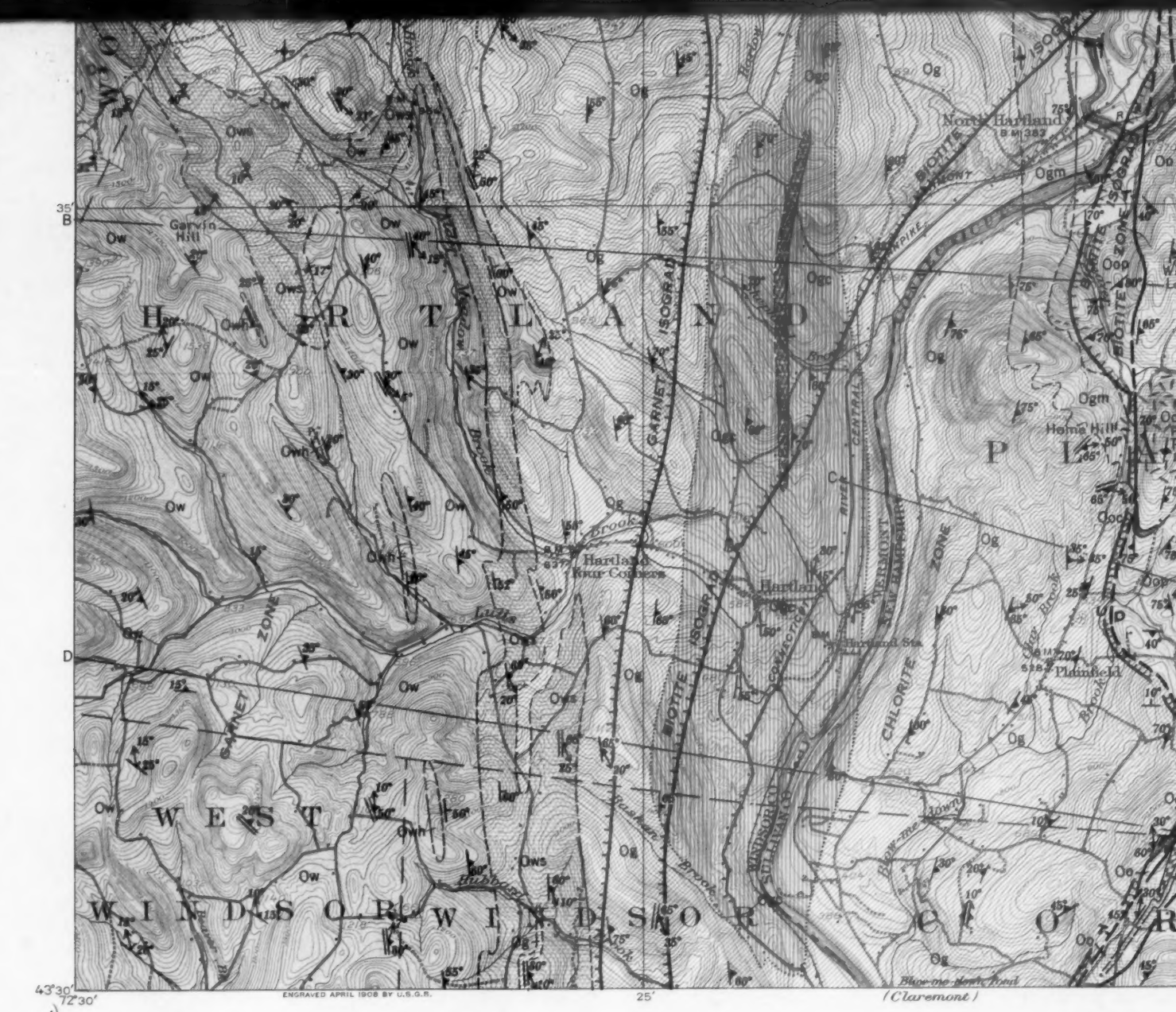
U D — High-angle fault; (U, upthrown side; D, downthrown side)

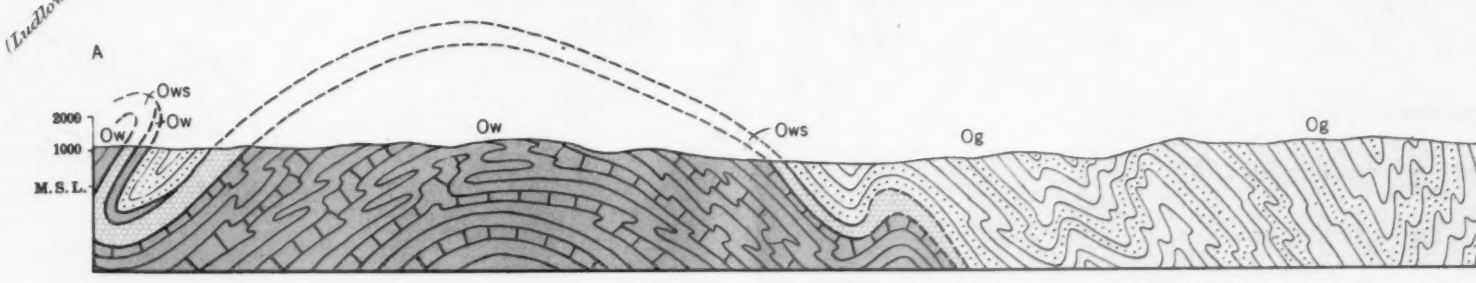
ANGLE, VERMONT AND NEW HAMPSHIRE

Geology mapped 1947-49 by J. B. Lyons, assisted by H. C. Coulter, J. F. Murphy, and H. H. Woodard.

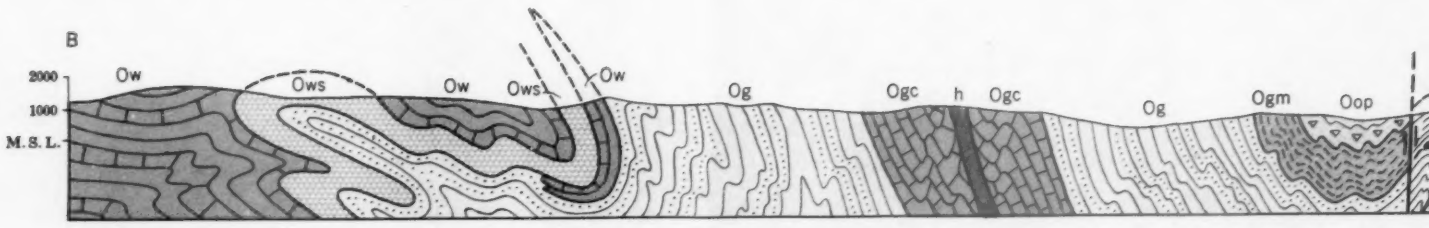
4 Miles

5 Kilometers

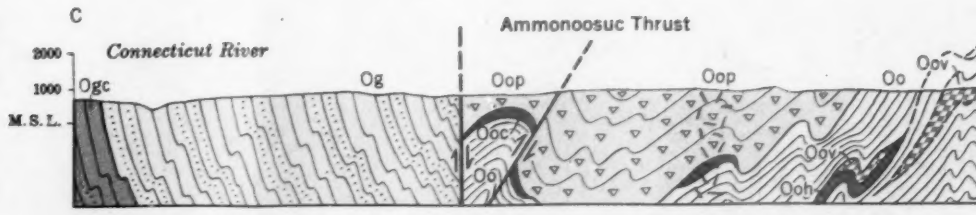




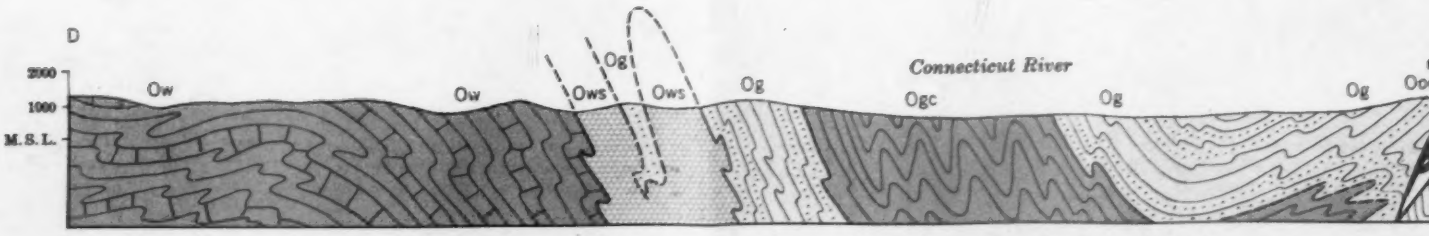
SECTION A-A'



SECTION B-B'



SECTION C-C'



SECTION D-D'

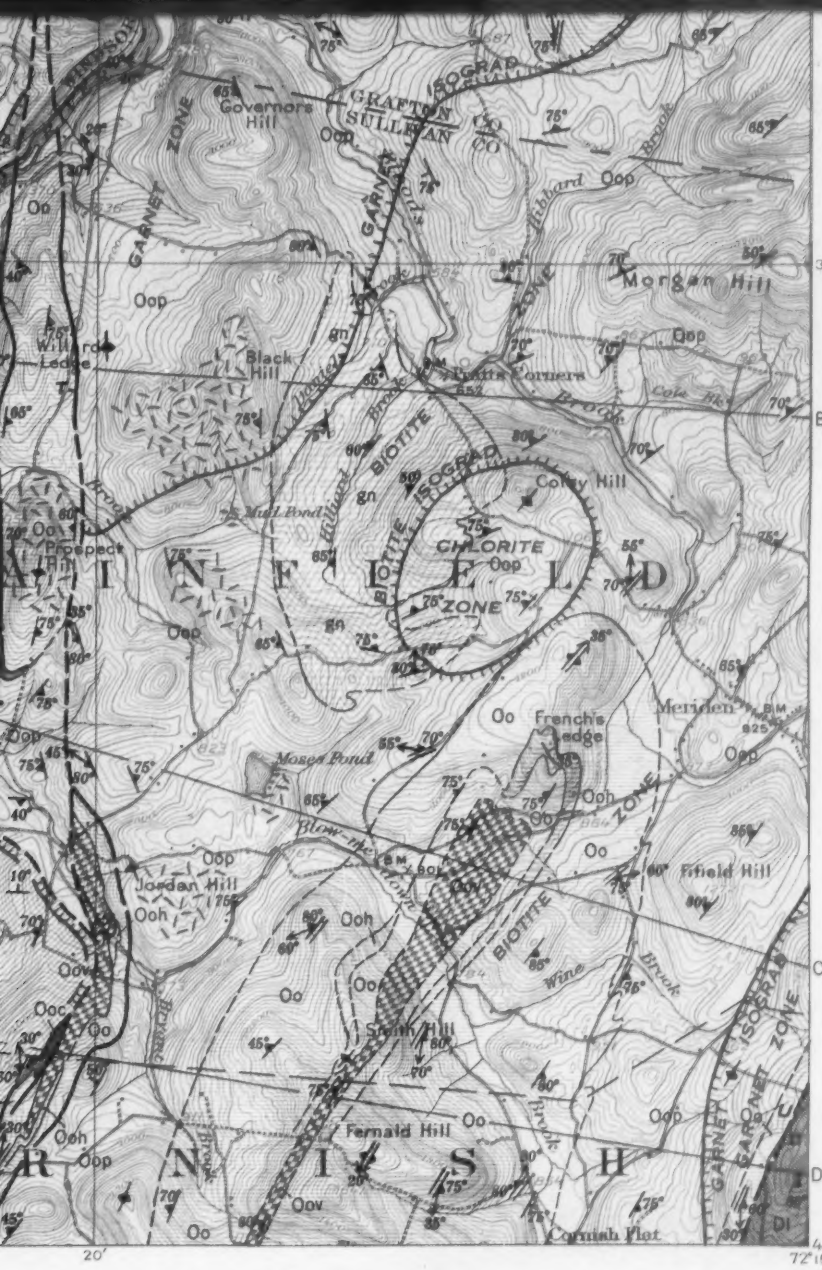
GEOLOGIC MAP AND STRUCTURE SECTIONS OF THE HANOVER QUADRA

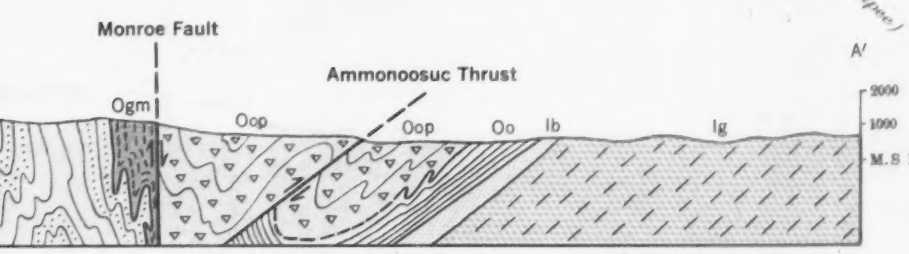
Topographic base by U. S. Geological Survey. Surveyed in 1905 and 1906

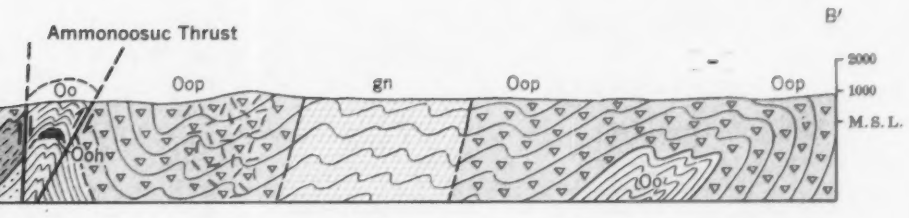
Scale 1/62500

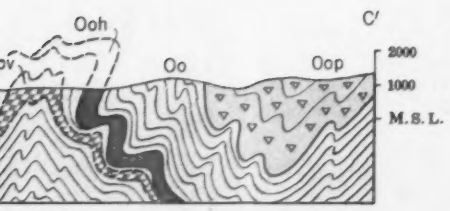
Contour interval 20 feet.

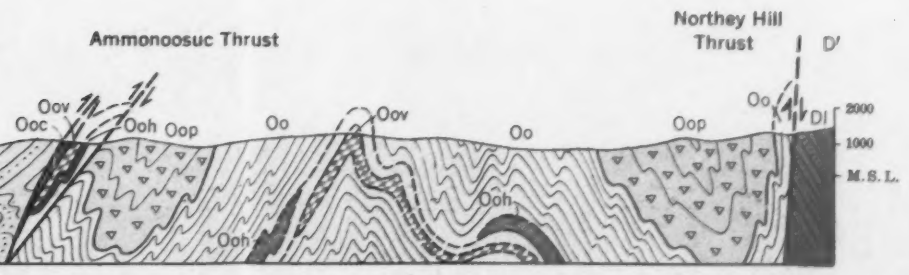
Datum is mean sea level.

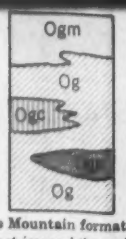






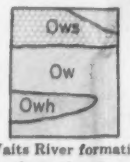






Gile Mountain formation

(West of the garnet isograd the rocks are tan to gray quartz-mica schist and micaceous quartzite, generally containing garnet and biotite porphyroblasts, Og. *East of the garnet isograd the same rocks,* Og, *are finer-grained, and lack garnet. The easternmost member of the formation, the Meetinghouse slate member,* Ogm, *is a black, thin-bedded phyllite and mica schist. Black calcite-quartz schist,* Ogc, *is widespread in the south-central part of the map area. Locally there are lentils of dark-green hornblende schist,* Ogn, *of probable volcanic origin.)*



Waits River formation

(Main part of the formation, Ow, *consists of brown-weathering calcite-quartz schist, with lesser amounts of mica schist and mica-quartz schist containing garnet porphyroblasts. Standing Pond amphibolite member,* Ows, *is chiefly a dark-green, garnetiferous, hornblende gneiss. Lentils of hornblende schist,* Owh, *are present locally.)*

NOTE: Albee and Ammonoosuc formations (Ordovician ?) and Clough and Fitch formations (Silurian) are cut out by the Northey Hill thrust where the Orfordville and Littleton formations are in juxtaposition. The Meetinghouse slate is, in part, the probable equivalent of the lower portion of the Orfordville formation.

METAMORPHIC ZONES

Metamorphic zones shown by isograds, hachured on the high-intensity side. Rocks east of the Ammonoosuc thrust in the southern portion of the area are in the biotite zone: those west of the thrust are in the chlorite zone.

CONTACTS

Accurate formation contact

Indefinite formation contact

Concealed or inferred contact

STRUCTURE SYMBOLS

25° — Strike and dip of bedding

50° — Strike and dip of overturned beds

30° — Strike and dip of foliation or schistosity

Strike of vertical foliation or schistosity

Horizontal foliation or schistosity

50° →40° — Foliation or schistosity, with strike and plunge of lineation

20° →30° — Strike and dip of axial plane of minor fold, with strike and plunge of fold axis

T — Thrust fault; (T, upper plate)

U D — High-angle fault; (U, upthrown side; D, downthrown side)

RANGLE, VERMONT AND NEW HAMPSHIRE

Geology mapped 1947-49 by J. B. Lyons, assisted by H. C. Coulter, J. F. Murphy, and H. H. Woodard.

2 4 Miles

5 Kilometers

Intro
Sc
Pr
Ackn
Strat
Ge
W
Gi
Or
Li
M
Plut
M
M

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 66, PP. 105-146, 12 FIGS., 3 PLS. JANUARY 1955

GEOLOGY OF THE HANOVER QUADRANGLE, NEW HAMPSHIRE-VERMONT

BY J. B. LYONS

ABSTRACT

The Hanover quadrangle, in west-central New Hampshire and east-central Vermont, is underlain predominantly by three metamorphosed eugeosynclinal formations, all of probable Ordovician age. The Waits River formation, with an exposed thickness of 4000 feet, and the Gile Mountain formation, with a thickness of 7500–8500 feet, are representative of the so-called Vermont sequence. In areas to the north and in part of the Hanover quadrangle, this sequence is separated from the New Hampshire sequence (here represented by the Orfordville formation, with a thickness of 6000 feet) by the Monroe fault. This fault apparently dies out in this quadrangle, and the two sequences are in sedimentary contact. Although the evidence is not unequivocal, it is concluded that Vermont rocks underlie the New Hampshire section and that the Gile Mountain formation is a probable equivalent to the lower part of the Orfordville formation and the upper part of the Waits River formation. A revision of the originally described stratigraphic sequence of the Orfordville formation is necessitated by the geologic relations of the Hanover area.

Two pre-metamorphic faults, the Monroe and Northey Hill, and the post-metamorphic Ammonoosuc thrust, disrupt the regional structures. The areal geology is dominated by two domal uplifts, the Lebanon dome and the Pomfret dome. The former is a syn-kinematic granite intrusive which has been forcibly injected from the northwest toward the southeast, overturning the beds south of the pluton. Considerable granitization has accompanied the intrusion of this massif. The Pomfret dome is a cleavage dome produced by a late-stage or second-deformation upward welling of previously deformed plastic schists. Granite or migmatite working its way upward appears to be the active tectonic element in the formation of the cleavage dome. Most of the minor structures of the Hanover area are tectonically related either to the Lebanon dome or to the Pomfret dome.

Regional metamorphism has affected all the rocks, and the intensity of metamorphism increases toward the domal structures. The latter are evidently the major heat sources in the metamorphic reconstitution of the interbedded sedimentary and volcanic rocks. Soaking of quartz, albite, and, to a lesser extent, tourmaline occurs in schists remote from Lebanon granite. Sedaclase tonalites, occasionally in *lit-par-lit* structures in these metasomatized rocks, possibly represent albitized granites, but their origin is conjectural.

CONTENTS

TEXT

	Page
Introduction	106
Scope of problem	106
Previous work	106
Acknowledgments	107
Stratigraphy	107
General statement	107
Waits River formation	109
Gile Mountain formation	111
Orfordville formation	113
Littleton formation	116
Missing formations	117
Plutonic rocks	117
Metamorphosed dikes and sills	117
Metasedimentary (?) dike	117
Diabase	118
Ultramafic rocks	118
Lebanon granite	118
Gneiss at White River Junction, Vermont	119
Gneiss east of Plainfield, New Hampshire	121
Structure	122
General statement	122
Pomfret dome	122
Recumbent folds of western part of quadrangle	125
Meriden anticline	127
Lebanon dome	127
Structure of area between Lebanon and Pomfret domes	130
Minor folds	130
Faults	132

Page
Cleavage 134
Lineation 135
Petrofabrics 136
Metamorphism 137
Introduction 137
Isograds and cause of metamorphism 138
Retrogressive metamorphism 139
Special features 139
Mineralogy of metamorphic rocks 140
Tectonic history 142
References cited 143

ILLUSTRATIONS

Figure Page
1. Index map 107
2. Columnar section 108
3. Sketch map of geology of a portion of Mt. Cube quadrangle 116
4. Sketch map of domal structures of northwestern New England 123
5. Sketch map and cross sections showing distribution and structure of Standing Pond amphibolite 124
6. Recumbent folds 126
7. Map of portion of north end of Lebanon pluton 128
8. Block diagram showing areal and structural relations of Lebanon granite 129
9. Sketch of folds in the Gile Mountain formation 131
10. Chevron folds 132
11. Isograd map of Hanover quadrangle 137
12. Plagioclase and plagioclase-epidote associations in rocks of Hanover area 140

Plate Facing page
1. Geologic map and cross sections of the Hanover quadrangle, New Hampshire-Vermont 105
2. Tectonic map of the Hanover quadrangle, New Hampshire-Vermont 122
3. Geologic map and cross sections of the Lebanon granite 128

TABLES

Table Page
1. Estimated modes of Waits River formation 110
2. Estimated modes of Gile Mountain formation 112
3. Estimated modes of Orfordville formation 115
4. Estimated modes of rocks of Lebanon pluton 119
5. Estimated modes of gneiss at White River Junction, Vermont 120
6. Estimated modes of gneiss east of Plainfield, New Hampshire 121
7. Estimated average chemical composition of gneiss east of Plainfield and of Post Pond volcanics 121

INTRODUCTION

Scope of Problem

During the years 1946–1950 six contiguous quadrangles in west-central New Hampshire and east-central Vermont (Fig. 1) were mapped concurrently by several geologists interested chiefly in the following problems: (1) stratigraphic and structural relations in the Connecticut Valley, and particularly the relative positions of the "New Hampshire" and "Vermont" sequences (White and Billings, 1951, p. 649); (2) the origin and mode of emplacement of two somewhat similar structural features—the Oliverian granitic domes of western New Hampshire and the cleavage domes of eastern Vermont; and (3) the nature and cause of the regional metamorphism. Information obtained in the other quadrangles has been used as a guide in this work, and a correlation of structural and stratigraphic units has been made, when possible, in the field. Statements of opinion on debatable points which will be set forth in this paper, however, do not necessarily represent the conclusions of the author's colleagues.

Previous Work

Partly because of the location of Dartmouth College in the northeastern ninth of the Hanover quadrangle, several papers have appeared describing the local geology. Bedrock studies have been made by C. T. Jackson (1844), E. Hitchcock (1861), C. H. Hitchcock (1874; 1877; 1878; 1908), J. W. Merritt (1913), and E. P. Kaiser (1938). Glacial features have been discussed by C. H. Hitchcock (1874; 1877;

1878; 1908), J. W. Goldthwait (1910; 1916), R. P. Lougee (1935), and J. W., L., and R. P. Goldthwait (1951). Recent mapping of the quadrangles to the north (Strafford quadrangle, C. H. Doll, 1944), northeast (Mt. Cube, J. B. Hadley, 1942; 1950), and east (Mascoma quadrangle, C. A. Chapman, 1939) makes it possible to correlate the Hanover area with others to the northeast (Littleton-Moosilauke, M. P. Billings, 1937) and northwest (Currier, L. W. and Jahns, R. H., 1941; W. M. Cady, 1950) where stratigraphic sequences have been worked out and where there is paleontologic evidence of the ages of the formations.

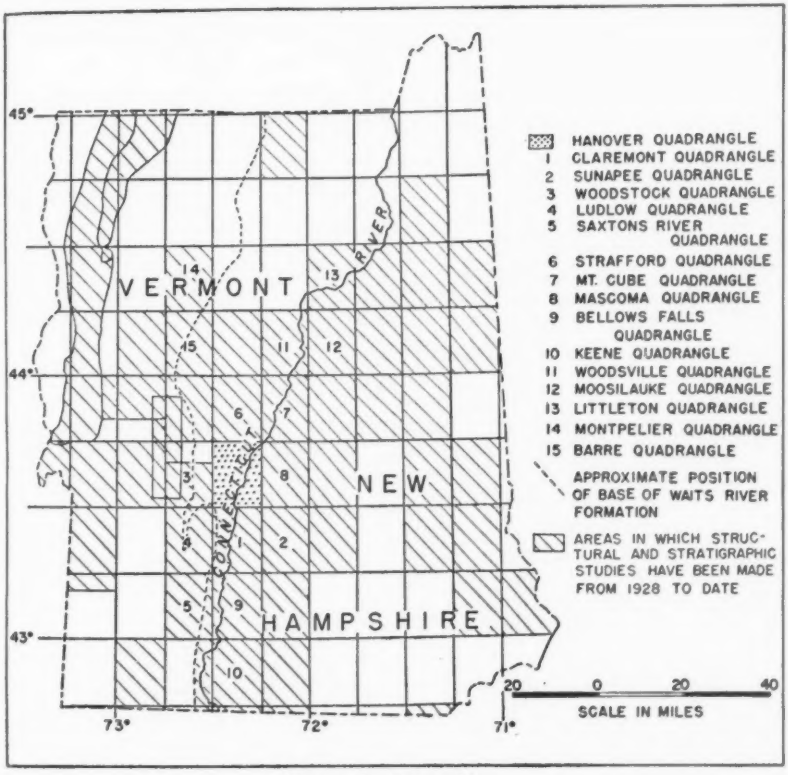


FIGURE 1.—INDEX MAP

ACKNOWLEDGMENTS

In addition to the assistance rendered by those men who have mapped the near-by areas, the writer has also profited from several discussions with Dr. W. M. Cady and Prof. M. P. Billings. The excellent field assistance of Messrs. H. C. Coulter, J. F. Murphy, and H. H. Woodard, all formerly of Dartmouth College, is gratefully acknowledged. Professors R. E. Stoiber, M. P. Billings, and A. H. McNair have critically read the manuscript. Field and laboratory expenses have been met largely by two generous grants from the Penrose Bequest of The Geological Society of America. The Committee on Research at Dartmouth College has defrayed part of the drafting expenses.

STRATIGRAPHY

General Statement

With the exception of a small area in the extreme southeastern part of this quadrangle, where metasediments of Devonian age are ex-

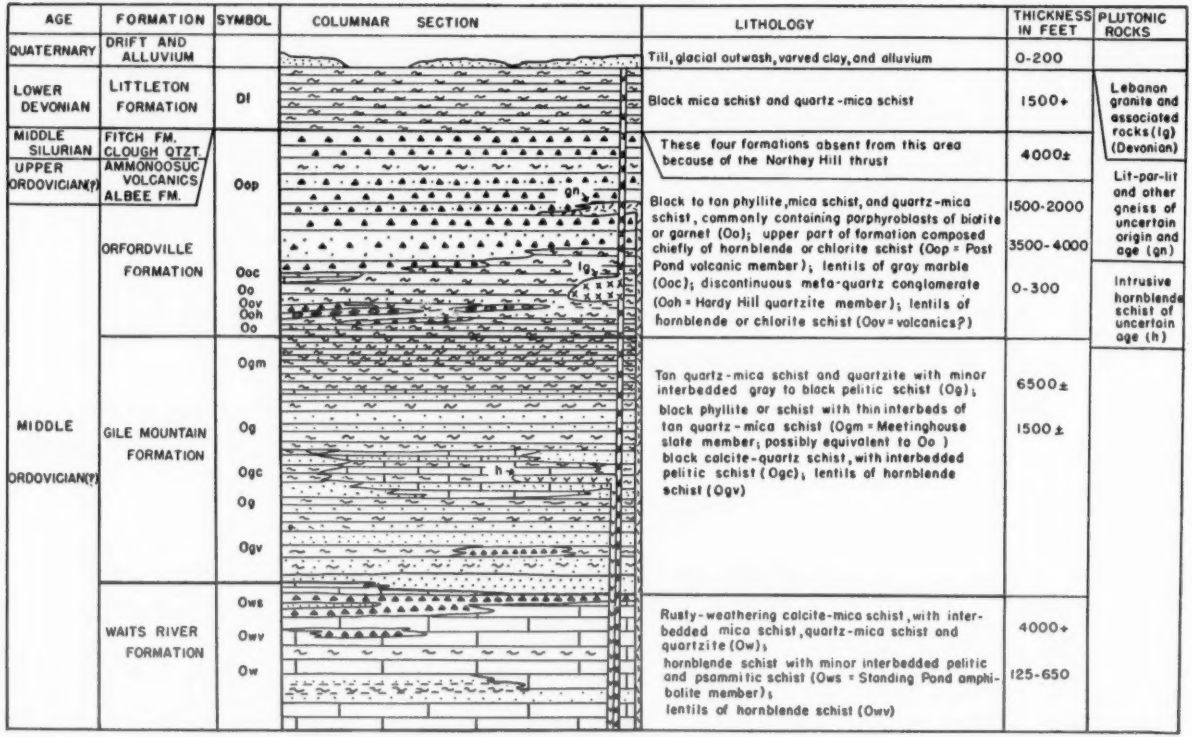

AGE
FORMATION
SYMBOL
COLUMNAR SECTION
LITHOLOGY
THICKNESS IN FEET
PLUTONIC ROCKS
QUATERNARY
DRIFT AND ALLUVIUM
Till, glacial outwash, varved clay, and alluvium
0-200
LOWER DEVONIAN
LITTLETON FORMATION
Dl
Black mica schist and quartz-mica schist
1500+
Lebanon granite and associated rocks (lg) (Devonian)
MIDDLE SILURIAN
FITCH FM.
CLOUGH QTZT.
UPPER ORDOVICIAN(?)
AMMONOOSUC VOLCANICS
ALBEE FM.
Oop
These four formations absent from this area because of the Northey Hill thrust
4000±
Lit-par-lit and other gneiss of uncertain origin and age (gn)
ORFORDVILLE FORMATION
Ooc
Oo
Oov
Ooh
Oo
Black to tan phyllite, mica schist, and quartz-mica schist, commonly containing porphyroblasts of biotite or garnet (Oo); upper part of formation composed chiefly of hornblende or chlorite schist (Oop = Post Pond volcanic member); lentils of gray marble (Ooc); discontinuous meta-quartz conglomerate (Ooh = Hardy Hill quartzite member); lentils of hornblende or chlorite schist (Oov = volcanics?)
1500-2000
3500-4000
0-300
Intrusive hornblende schist of uncertain age (h)
MIDDLE ORDOVICIAN(?)
GILE MOUNTAIN FORMATION
Ogm
Og
Ogc
Og
Ogv
Tan quartz-mica schist and quartzite with minor interbedded gray to black pelitic schist (Og); black phyllite or schist with thin interbeds of tan quartz-mica schist (Ogm = Meetinghouse slate member; possibly equivalent to Oo) black calcite-quartz schist, with interbedded pelitic schist (Ogc); lentils of hornblende schist (Ogv)
6500±
1500±
WAITS RIVER FORMATION
Ows
Owv
Ow
Rusty-weathering calcite-mica schist, with interbedded mica schist, quartz-mica schist and quartzite (Ow); hornblende schist with minor interbedded pelitic and psammitic schist (Ows = Standing Pond amphibolite member); lentils of hornblende schist (Owv)
4000+
125-650

FIGURE 2.—COLUMNAR SECTION
Rocks of Hanover quadrangle.

posed, three major stratigraphic units constitute the structural framework of the region (Pl. 1). These formations, of probable Ordovician age, represent a typical metamorphosed eugeosynclinal sequence consisting chiefly of argillaceous, arenaceous, and calcareous metasediments and metamorphosed volcanics. Their aggregate thickness is approximately 18,000 feet, although this figure may be considerably in error. As is true throughout most of New Hampshire and Vermont, rocks of the lowest intensity of metamorphism lie at or just west of the Connecticut River; isograds rise progressively toward the granitic plutons on the east and the cleavage domes (White and Jahns, 1950, p. 209) on the west. The general stratigraphic relations of the formations in the Hanover area are shown in the columnar section (Fig. 2).

Waits River Formation

DISTRIBUTION: The Waits River formation (Currier and Jahns, 1941, p. 1491) is exposed in an elliptical area underlying most of the northwestern ninth of this quadrangle, and in a southeasterly trending belt in the west-central and southwest ninths. The best exposures are along or just west of the White River Valley (Pl. 1).

LITHOLOGY: Most of the Waits River formation consists of the following rocks, in order of decreasing importance: (1) tan calcite-quartz-plagioclase schists and marbles which weather into distinctive brown "rottenstones", (2) gray mica schists, and (3) tan quartzites and quartz-mica schists. Transgressing from near the top of the Waits River formation into the overlying Gile Mountain formation is the Standing Pond amphibolite, first mapped by Doll (1944, p. 17) as an uppermost member of the Waits River formation in the Strafford quadrangle. Thin discontinuous bands of hornblende schist within the Waits River are indicated on Plate 1; these are believed to represent for the most part metasediments of volcanic derivation.

Individual beds within the formation average from 4 to 10 feet in thickness, but a few beds in the southern part of the area attain thicknesses of 30 feet or more. A marked increase in the thickness and abundance of the arenaceous metasediments of the Waits River is evident toward the south. The Gile Mountain formation in the same area becomes increasingly calcareous, so that on a purely lithologic basis the two units cannot be readily separated. These lithologic changes are in keeping with the fact that the Waits River formation thickens and becomes more calcareous toward the northwest (White and Jahns, 1950); they also explain why the Gile Mountain formation has not been mapped or recognized to the south (Bellows Falls quadrangle, F. C. Kruger, 1946; Keene-Brattleboro area, G. E. Moore, 1949).

Table 1 summarizes the salient features of the mineralogy of the formation.

The Standing Pond amphibolite, which ranges from 125 to 650 feet in thickness, contains a number of interstratified metasedimentary lentils. Garnet porphyroblasts near the upper and lower contacts commonly attain a diameter of more than an inch, lending a distinctive appearance to outcrops of this important stratigraphic marker. Despite its ferromagnesian composition, the presence of 10 per cent of more of quartz in all the amphibolite and the intergradation of amphibolite and metasediments make it probable that the unit was originally neither a flow nor an intrusion, but a sedimentary rock composed of volcanic debris. It is a time-stratigraphic unit which transgresses lithologic boundaries. Paleogeologically it is important because it proves that the Waits River and Gile Mountain formations were deposited without a break in sedimentation; they must, therefore, be essentially of the same geologic age.

THICKNESS: Because the base of the Waits River formation is not exposed in the Hanover area, reliable estimates of its thickness are not possible. Structure sections drawn from the known field and map data would require a minimum thickness of approximately 4000 feet. Estimates made elsewhere are highly variable. Chang (1950, p. 11) has concluded that the Waits River formation within the Woodstock quadrangle (directly west of the Hanover quadrangle) is only 2500 feet thick. Moore (1949, p. 1619) has arrived at a minimum value of 4000–5000 feet in southern Vermont. White and Jahns (1950, p. 191) set a minimum figure of 10,000 feet in central Vermont. The combined thick-

TABLE 1.—ESTIMATED MODES OF WAITS RIVER FORMATION

Rock	1	2	3	4	5	6	7	8	9
No. of sections	6	1	2	2	2	13	3	1	2
Quartz	25	5	30	66	12	20	.....	83	....
Plagioclase	10	1	15	15	33	20	44	....	47
Calcite	50	60	.....	1	4	2	4	....	t
Ankerite	t	20	.....	....	3	.....	.....	....	....
Muscovite	10	7	25	4	....	t	....	....	....
Biotite	3	2	15	5	15	15	4	....	32
Chlorite	1	1	2	2	8	3	2	t	3
Hornblende	....	....	.....	....	20	31	40	3	4
Actinolite	....	....	.....	....	....	t	.....	2	....
Clinozosite	1	....	.....	....	1	4	5	....	1
Garnet	....	....	5	....	....	t	.....	10	8
Staurolite	....	....	1	2	....	.....	.....	....	....
Kyanite	....	....	3	....	....	.....	.....	....	....
Apatite	t	....	.....	1	1	2	1	2	1
Zircon	t	....	.....	x	t	t	.....	....	....
Allanite (?)	x	....	1	1	....	x	.....	....	....
Tourmaline	x	....	.....	1	....	.....	.....	....	....
Sphene	t	....	.....	....	....	x	.....	....	....
Pyrite	x	....	1	....	....	t	.....	....	....
Magnetite	t	3	2	2	3	3	.....	t	4
Ilmenite	....	....	.....	....	....	t	.....	....	....
Carbon	....	1	t	....	....	.....	.....	....	....
% An in plag.	25	25	27–45	20	27	20–32	23–38	....	25

x = present in 50% or more of thin sections
t = present in less than 50% of thin sections, or in very minor amounts in some sections

1. Calcite-quartz schist
2. Marble
3. Mica schist
4. Quartz-mica schist
5. Hornblende schist
6. Standing Pond amphibolite; hornblende schist
7. Standing Pond amphibolite; amphibolite
8. Standing Pond amphibolite; garnetiferous quartzite
9. Standing Pond amphibolite; feldspathic schist

ness of the Ayers Cliff and Barton River formations of northern Vermont (equivalents of the Waits River) is approximately 13,000 feet (Doll, 1951, p. 23, 26).

CORRELATION: The Waits River formation of the Hanover quadrangle may be traced directly to the outcrops at Waits River, Vermont, in the East Barre quadrangle, where C. H. Richardson (1906, p. 86) first named the unit the Waits River limestone. In the Strafford quadrangle (immediately north of the Hanover quadrangle) Doll (1944, p. 16) has applied the name Memphremagog formation to the same beds. In northern Vermont Doll (1951, p. 22) has recently subdivided the Waits River equivalents into three units: the Ayers Cliff, Barton River, and Westmore formations. The Westmore may correlate with the Gile Mountain formation. Canadian equivalents of the Waits River formation are the Tomifobia slates (Clark, 1934) which Doll (1951, p. 23) correlates with the Ayers Cliff (lower Waits River) formation. The Tomifobia slates Clark (Cady, 1950, p. 489) believes are the equivalents of the St. Francis

series (Cooke, 1937, p. 37–42) which crops out east of Thetford Mines, Quebec.

AGE: Because no fossils have been found in the Hanover quadrangle, no direct contribution can be made to the problem of the age of the Waits River formation. The existing fossil evidence is controversial. Fossil localities within the Waits River or its equivalents at Montpelier, Vermont (Cady, 1950), and in Quebec (Clark, 1934; Cooke, 1937) have yielded, respectively, corals and graptolites which have been assigned to the Middle or Upper Ordovician. Moreover the Waits River formation of central Vermont is underlain by the unfossiliferous Northfield slate, and this, in turn (with slight unconformity), by the fossiliferous Middle Ordovician Shaw Mountain formation (Currier and Jahns, 1941, p. 1501). All these data seemingly establish the age of the Waits River as Middle or Upper Ordovician.

Doll (1943a, p. 60–62), however, has collected from the Westmore formation at Westmore, Vermont, crinoids, cystoids, and gastropods which have been assigned to the Middle Silurian or Lower Devonian. He has also (1943b, p. 676) described a brachiopod from the Gile Mountain of the Strafford quadrangle which has likewise been assigned to the Lower Devonian. The Westmore and Gile Mountain formations, as has been indicated previously, are probably equivalent stratigraphically.

Structural evidence throughout most of Vermont would, in the writer's opinion, favor an Ordovician rather than a Devonian or Silurian age for the Waits River formation. The effects of the Taconic orogeny should be expressed in a strong unconformity somewhere in the stratigraphic column. Such an unconformity has been described, for example, at the base of the Silurian in western New Hampshire (Billings, 1937) and has been noted at several places in southern Quebec. Except for the small unconformity at the base of the Northfield slate, the stratigraphic units below the Waits River formation and above the Precambrian rocks of the core of the Green Mountains are apparently conformable (Thompson, Rosenfeld, and Cady, oral communication). Doll (1951) has argued that the Irasburg conglomerate of northern Vermont, which separates the Ayers Cliff and Barton River formations, is tectonically significant. This conglomerate, however, is not areally widespread nor is it persistent in central and southern Vermont. The Irasburg conglomerate would lie within the apparently homogeneous Waits River formation at the latitude of the Hanover quadrangle.

Gile Mountain Formation

DISTRIBUTION: The Gile Mountain formation (Doll, 1944, p. 18) crops out in a belt 5–8 miles wide extending in a northerly direction through the center of the Hanover quadrangle. It is the most widespread formation mapped and underlies an area of approximately 100 square miles. Excellent exposures of the entire formation from its basal quartzites to its uppermost member, the Meetinghouse slate (Doll, 1944, p. 19), occur along the White River, between the villages of West Hartford and Hartford, Vermont (Pl. 1).

LITHOLOGY: The major rocks constituting the Gile Mountain formation are tan quartz-biotite schist and interbedded gray mica schist and phyllite. Toward the west, near its contact with the Waits River formation, the beds are arenaceous, and quartzites 20–30 feet thick are common; toward the east there is an increase in the amount of argillaceous material and a thinning of individual beds. The easternmost and, according to Doll (1944), the uppermost part of the formation is the Meetinghouse slate member. This member grades into the Gile Mountain formation on the west and has been arbitrarily separated from it in this quadrangle by the following criteria: (1) dark phyllite constitutes 50 per cent or more of the observable outcrops, (2) individual beds average less than 4 feet thick, (3) "pin-stripe" schists, consisting of quartzose layers half a cm in average width separated by chlorite-sericite laminae, are characteristic of the easternmost few hundreds of feet of outcrop of this member.

A locally developed facies of interbedded black quartz-calcite schist and phyllite underlies a large portion of the south-central ninth of the quadrangle (Ogc on Pl. 1); these rocks are similar to the calcareous portions of the Gile Mountain formation mapped by Hadley (1950, Pl. 2) in the northwestern part of the Mt. Cube

quadrangle. Lenses of amphibolite and hornblende schist, some of them of intrusive origin, occur sporadically. As has been pointed out, the southwesternmost exposures of the Gile Mountain formation are calcareous, and the formation is difficult to separate lithologically from the Waits River formation. The Meetinghouse slate, traced southward along its strike, becomes increasingly arenaceous; in the north it is a black phyllite or mica schist, but in the south tan quartz-mica schist predominates.

Mineralogic characteristics of the Waits River formation are indicated in Table 2.

THICKNESS: Doll (1944, p. 19) has estimated a total thickness for the Gile Mountain formation of 7600 feet, of which he assigns 1100 feet to the Meetinghouse slate member. These estimates are probably valid for this formation within the Hanover quadrangle, although our ignorance of its detailed structures makes any such figure susceptible to large errors.

CORRELATION: The Gile Mountain formation and the Meetinghouse slate member are directly traceable from this quadrangle to their type localities in the Strafford quadrangle (Doll, 1944, p. 18–19). As will be discussed more fully following the description of the Orfordville formation, it is tentatively concluded in this paper that the Meetinghouse slate and the lower portion of the Orfordville are stratigraphically

TABLE 2.—ESTIMATED MODES OF GILE MOUNTAIN FORMATION

Rock	1	2	3	4	5	6	7	8
No. of sections	2	1	7	2	2	1	1	2
Quartz	87	50	62	20	18	3	60	47
Plagioclase	.....	5	8	40	.....	60	10	15
Calcite	.....	.....	t	3	72	5	.....	5
Muscovite	4	2	7	5	7	.....	18	20
Biotite	5	15	13	10	.....	23	2	2
Chlorite	.....	4	2	3	.....	2	5	7
Hornblende	.....	.....	.....	11	.....	.....	.....	.....
Clinozoisite	.....	.....	.....	1	.....	3	.....	.....
Garnet	.....	7	6	5	.....	.....	.....	.....
Staurolite	.....	3	.....	.....	.....	.....	.....	.....
Kyanite	.....	10	.....	.....	.....	.....	.....	.....
Apatite	x	x	x	t	.....	1	1	1
Zircon	x	x	x	t	.....	.....	t	.....
Allanite (?)	.....	.....	t	t	t	.....	.....	.....
Tourmaline	.....	x	t	1	.....	.....	1	x
Sphene	.....	.....	.....	.....	.....	.....	.....	1
Magnetite	4	3	2	1	1	3	3	2
Carbon	.....	.....	t	.....	2	.....	.....	.....
% An in plag.	.....	21	21–28	30	.....	9	11	4–6

x = present in 50% or more of thin sections
t = present in less than 50% of thin sections; or in very minor amounts in some sections

1. Quartzite
2. Kyanite-mica schist
3. Mica-quartz schist
4. Hornblende schist
5. Calcite-quartz schist
6. Biotite-feldspar schist
7. Phyllite
8. "Pin-stripe" mica-quartz schist; Meetinghouse slate member

equivalent. These two units have been differentiated, however, on the geologic map (Pl. 1) because of some uncertainties in this correlation.

AGE: The stratigraphic relations of the Standing Pond amphibolite make it clear that the Waits River and Gile Mountain formations grade into one another and are, therefore, partially contemporaneous. The comments set forth on the age of the Waits River formation are equally applicable to the Gile Mountain formation, and its age is probably Middle or Upper Ordovician.

Orfordville Formation

GENERAL STATEMENT: As mapped in the Hanover quadrangle the Orfordville formation consists, from bottom to top, of the four following members: Orfordville schists, Hardy Hill quartzite, Orfordville schists, and Post Pond volcanics. This stratigraphic succession differs from that in Hadley's (1942, p. 119–124) original description of the Orfordville formation. Reasons for the stratigraphic revision will be set forth at the end of the description of this formation.

DISTRIBUTION: The Orfordville formation underlies most of the eastern third of the Hanover quadrangle. Middle-grade rocks of the Post Pond volcanic member are well exposed in Bloods Brook, which flows northwesterly from Meriden, N. H., into the Connecticut River. Low-grade representatives of the Post Pond member crop out in a series of cliffs extending westerly from the Connecticut River through the village of Norwich, Vermont. Schists of the Orfordville formation form prominent outcrops on Farnum Hill and Storrs Hill, south of Lebanon, N. H. Hardy Hill quartzite holds up French's Ledge southwest of the village of Meriden, N. H.

LITHOLOGY: Depending upon the degree of metamorphism, the Post Pond volcanics are either hornblende or chlorite schists, with minor interstratified metasedimentary lenses. Their volcanic nature is locally indicated by poorly preserved pillow structure and by relict agglomerates and amygdules. Numerous dikes and sills of hornblende schist and amphibolite intrude the Post Pond; some of these possibly represent fissure feeders for some of the extrusions. Some of the amphibolites and chlorite schists are quartz free and have modal compositions which indicate that they were originally basaltic. Most of the schists, however, are quartzose. The average chemical composition of the entire member (estimated from its modes) is that of a basalt to which approximately 20 per cent SiO_2 has been added. From this it seems probable that relatively little of the volcanic material was deposited as such; most of it was reworked and intermixed with quartz. A minor proportion of the Post Pond volcanics is composed of rocks which were once felsic flows, some of them probably having been quartz keratophyres prior to metamorphism (Table 3).

Schists and phyllites in the Orfordville formation have a variable nature depending upon their geographic and geologic position. In the northeastern part of the area the commonest lithologic type is a black garnetiferous schist or phyllite; tan garnetiferous quartz-mica schists and green garnetiferous hornblende-biotite schists are minor members. Toward the southeast the formation becomes increasingly arenaceous. Quartz-mica schist is the predominant rock type, and the resemblance to the arenaceous southern exposures of the Meetinghouse slate is striking.

In the Hanover quadrangle, the horizon between the schists of the Orfordville formation and their overlying Post Pond member is locally marked either by calcareous lenses in the lower part of the Post Pond or by lentils of gray marble or recrystallized limestone. The latter are conspicuous on outcrops at the south end of Prospect Hill in Plainfield, N. H. (Pl. 1, Ooc), and on a hill 1½ miles southeast of that village. In both localities the lentils are at least 100 feet thick.

The Hardy Hill quartzite is a meta-quartz conglomerate and quartzite characterized by abundant deformed pebbles of coarsely granular white quartz and quartzite. Maximum average diameter of the pebbles ranges up to 4 inches. Their provenance is unknown, and the origin and genetic significance of the Hardy Hill quartzite as a whole is not well understood. Krumbein and Sloss (1951, p. 368) ascribe quartzose conglomerates such as the Hardy Hill to shelf environments tempo-

rarily developed during the evolution of a geosyncline.

Peripheral to the "border gneiss" of the Lebanon granite is a group of quartzo-feldspathic gneisses and schists (Pl. 1, Oog; Table 3, columns 19–21) which have gradational contacts both with the border gneiss and with the surrounding Orfordville formation. These rocks are of such a distinctive appearance that, although they are a part of the Orfordville formation, they have been mapped separately. Several lines of evidence indicate that they have undergone thorough metasomatism by feldspathizing emanations from the Lebanon granite. The stratigraphic position of these gneisses and schists is somewhat uncertain. Chapman (1939, Pl. 1), basing his correlation on their feldspathic nature and the presence of thin lentils of hornblende and chlorite schist, grouped them with the Post Pond volcanics. This correlation is untenable not only because of the striking petrographic differences of the two units but also because it involves structural interpretations which are not borne out by the field and map data. One of the quartzose beds on the eastern side of the metasomatized rocks (Table 3, column 20) possesses a relict conglomeratic structure and has been interpreted by the writer (Pl. 3) as a probable equivalent of the Hardy Hill quartzite. This rock crops out on Signal and Hayes Hills, Etna, on the eastern side of an area of quartz-kyanite schists described by Bannerman (1941) as being potentially exploitable.

Generalized mineralogical compositions of the rocks of the Orfordville formation are indicated in Table 3.

THICKNESS: The thickness of the Orfordville formation is highly variable because of changes in the thickness of the Post Pond member. In the Hanover quadrangle the latter is estimated to have a thickness of 3500–4000 feet, and the sedimentary schists of the formation approximately 2000 feet thick. Maximum thickness of the Hardy Hill quartzite is 350 feet, but this member thins and pinches out along the strike. Hadley (1942) estimated 5000 feet as the total thickness of the formation in the Mt. Cube quadrangle.

AGE: No fossils have been found in the Orfordville formation in New Hampshire. The author would concur with Hadley (1942, p. 124) that its age is probably Middle or Upper Ordovician.

CORRELATION: The Orfordville formation can be traced directly into the Mt. Cube quadrangle where it was first described by Hadley (1942, p. 119–124), and most of its members were named. Chapman named (1939, p. 132) the Hardy Hill quartzite from outcrops in the Mascoma quadrangle.

REVISION OF ORFORDVILLE STRATIGRAPHY: In Hadley's (1942, p. 119) original description of the Orfordville formation the following subdivisions were adopted (from top to bottom):

Sunday Mountain volcanic member
Dark schists
Hardy Hill quartzite member
Dark schists
Post Pond volcanic member
Dark schists

The Post Pond volcanic member of the Orfordville formation is directly traceable from the Mount Cube into the Hanover quadrangle. Southwest of the village of Meriden (Pl. 1) is a well-defined anticline in which are exposed Orfordville schists stratigraphically lower than the Post Pond volcanics. Within these schists is a meta-conglomerate identical in thickness and lithology with the Hardy Hill quartzite. This meta-conglomerate is separated from Post Pond volcanics by an outcrop width of Orfordville schists identical with that which separates near-by outcrops of the Hardy Hill and Post Pond members in the Mascoma quadrangle, a few miles to the east (Chapman, 1939, Pl. 1).

Inasmuch as the conglomerate in the Meriden anticline is stratigraphically below the Post Pond volcanics, this would mean that the Orfordville formation could be neatly subdivided into two major units, each consisting (from bottom to top) of schist, meta-conglomerate, schist, and meta-volcanics. The identification of the conglomerate in the Meriden anticline with the Hardy Hill quartzite eliminates the necessity of this highly improbable repetition and greatly simplifies the interpretation of the areal structure and stratigraphy; it also implies the essential equivalence of the lithologically similar Post Pond and Sunday Mountain volcanic members of the Orfordville formation.

Two additional lines of evidence buttress the

TABLE 3.—ESTIMATED MODES OF ORFORDVILLE FORMATION

TABLE 3.—ESTIMATED MODES OF ORFORDVILLE FORMATION

Rock	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21
No. of sections	6	7	3	1	6	1	1	1	4	4	1	1	5	1	4	8	1	1	7	3	8
Quartz	18	20	55	10	50	20	35	50	50	24	48	20	60	33	20	50	10	90	40	75	31
Plagioclase	30	30	20	..	25	35	51	10	10	30	..	30	..	55	45	20	..	..	25	..	30
Calcite	5	t	..	83	x	20	..	..	t	....	..	..	t	..	5	x	85	..	t	..	3
Ankerite	...	..	..	..	..	..	..	..	....	....	..	..	t	..	5	..	..	..	..	..	...
Muscovite	t	..	..	..	5	10	5	10	15	....	..	..	15	2	7	15	2	9	3	8	2
Biotite	t	..	t	..	x	..	2	15	10	10	..	3	10	1	2	5	..	1	10	t	12
Chlorite	6	8	10	5	15	15	5	5	5	....	..	2	10	5	16	10	..	..	t	..	2
Hornblende	35	33	10	..	..	..	..	..	....	30	5	20	..	..	....	..	..	..	t	..	8
Actinolite	...	..	..	..	..	..	..	..	....	....	..	..	..	2	....	..	..	..	..	..	...
Clinozoisite	3	6	2	..	4	t	..	..	2	4	45	25	..	t	....	..	..	..	6	..	9
Garnet	t	..	t	..	t	..	..	2	7	2	2	..	5	..	....	..	..	..	t	..	2
Staurolite	...	..	..	..	..	..	..	5	....	....	..	..	..	..	....	..	..	..	..	..	...
Kyanite	...	..	..	..	..	..	..	..	....	....	..	..	..	..	....	..	..	..	..	15	...
Microcline	...	..	t	..	..	..	..	..	....	....	..	..	..	..	....	..	..	..	15	..	...
Apatite	1	1	1	..	x	..	t	1	t	t	..	t	t	t	x	x	..	..	1	..	1
Zircon	...	..	..	..	t	..	..	t	...	t	..	..	t	..	....	t	..	..	..	t	t
Allanite (?)	...	..	..	..	..	..	..	t	t	....	..	..	..	..	....	..	..	..	t	t	t
Tourmaline	...	..	..	..	t	..	..	..	x	....	..	..	t	..	t	t	..	x	t	..	t
Sphene	1	2	..	2	t	..	..	..	....	....	..	..	..	..	....	..	..	..	..	..	t
Pyrite	...	x	..	..	..	..	..	..	....	....	..	t	..	..	x	t	1	..	..	2	...
Magnetite	x	x	2	..	1	t	2	2	1	t	..	..	t	2	x	x	..	x	..	..	t
Ilmenite	1	x	..	..	t	..	..	..	....	....	..	..	..	..	....	..	..	..	..	..	...
Carbon	...	..	..	..	..	..	..	..	t	....	..	..	..	..	....	t	2	..	..	..	t
% An in plag.	23–30	6–10	9–16	..	5–11	9	7	28	20–50	38–44	..	10	..	6	32–40	5–11	..	..	5–30	..	10–60

x = present in 50% or more of thin sections
t = present in less than 50% of thin sections; or in very minor amounts in some sections

1. Hornblende schist; Post Pond volcanics; amphibolite facies
2. Hornblende schist; Post Pond volcanics; albite-epidote-amphibolite facies
3. Hornblende-chlorite schist; Post Pond volcanics; albite-epidote-amphibolite facies
4. Marble; Post Pond volcanics; albite-epidote-amphibolite facies
5. Chlorite schist; Post Pond volcanics; greenschist facies
6. Chlorite schist; Post Pond volcanics; greenschist facies
7. Quartz-feldspar schist; Post Pond volcanics; greenschist facies; relict agglomeratic structure; probably originally a quartz keratophyre (sodaclase dacite)
8. Staurolite schist; amphibolite facies
9. Mica schist; amphibolite facies
10. Hornblende schist; amphibolite facies
11. Epidote granulite; amphibolite facies
12. Hornblende schist; albite-epidote-amphibolite facies
13. Mica schist; albite-epidote-amphibolite facies
14. Feldspathic schist; albite-epidote-amphibolite facies; originally a quartz keratophyre (?)
15. Feldspathic chlorite schist; albite-epidote-amphibolite facies (disequilibrium); metamorphosed volcanic?
16. Mica-quartz schist or phyllite; greenschist facies
17. Recrystallized limestone; greenschist facies
18. Hardy Hill quartzite; greenschist facies
19. Biotite gneiss; amphibolite facies; metasomatized Orfordville
20. Kyanite schist; amphibolite facies; metasomatized Hardy Hill
21. Hornblende-biotite gneiss; amphibolite facies; metasomatized Orfordville

proposed correlation. Graded bedding in the meta-conglomerate of the Meriden anticline is a reliable structural criterion for top-and-bottom relations and indicates that the Post Pond is that the elliptical area is a dome whose eastern limb has been strongly overturned and merged into a northward-plunging syncline which lies immediately east of the dome itself.

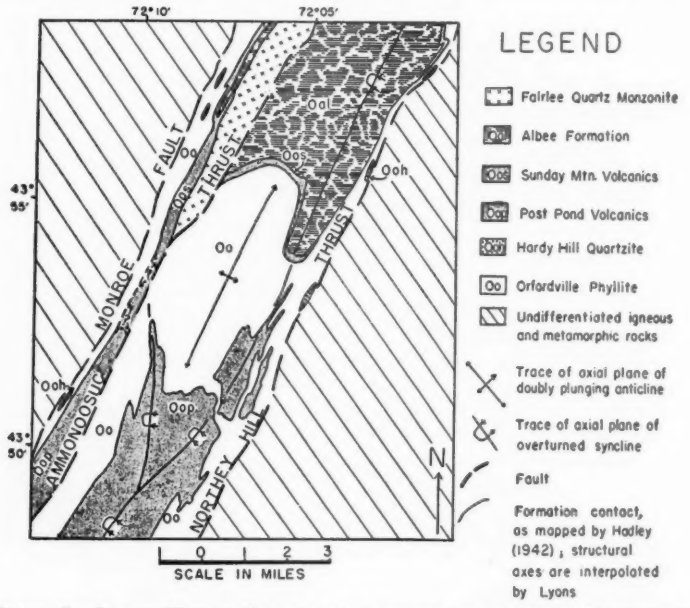


FIGURE 3.—SKETCH MAP OF GEOLOGY OF A PORTION OF MT. CUBE QUADRANGLE

volcanics overlie this unit. Graded bedding in the Hardy Hill quartzite at its type locality in the Mascoma quadrangle likewise indicates that the Post Pond volcanics are stratigraphically above that quartzite. Hence follows the probability that the two quartzites are equivalent.

A sketch map of part of the Mt. Cube quadrangle (Hadley, 1942, Pl. 1) is shown in Figure 3. The original interpretation of the stratigraphy of the Orfordville formation was based upon the geology of the elliptical area outlined on the north by the Sunday Mountain volcanics and on the west, east, and south by the Post Pond volcanics. This was described (Hadley, 1942, p. 157) as "an anticline (on the north) . . . replaced southward by a syncline." Without going into the complicated detailed geology of the area shown it may be stated that another interpretation, and one that is in harmony with the local and areal structure and with the proposed revision of the Orfordville stratigraphy, Post Pond volcanics on the northern extension of this syncline (the Salmon Hole Brook syncline) are approximately 400 feet stratigraphically below the Sunday Mountain volcanics. The wedging out of the Post Pond volcanics toward the north and their interdigitation with the Orfordville schists means that this relatively small difference in the stratigraphic positions of the Post Pond and Sunday Mountain members is inconsequential.

Littleton Formation

A series of black schists cropping out in the southeastern corner of the Hanover quadrangle is continuous with a belt which has been traced by Chapman (1939, Pl. 1) and Hadley (1942, Pl. 1) through the Mascoma and Mt. Cube quadrangles into outcrops of the Littleton formation in the Littleton-Moosilauke area (Billings, 1937, Pl. 1). According to Billings

(1937, p. 494) these rocks are of Devonian age and are separated from the rocks to the west by the Northey Hill thrust.

The Littleton formation in the southeastern portion of the Hanover quadrangle closely resembles the near-by outcrops of the Orfordville formation, except that the Littleton possesses a distinctive "pin-stripe" structure. The rock itself is a quartz-mica schist and has no unique mineralogical features. Its width of outcrop within this area is less than 2000 feet. Thickness cannot be estimated.

Missing Formations

The juxtaposition of the Orfordville and Littleton formations means that, in the latitude of the Hanover quadrangle, the Ordovician (?) Albee formation and Ammonoosuc volcanics and the Silurian Clough and Fitch formations have been eliminated by the Northey Hill thrust. For descriptions of these formations the reader is referred to Billings (1937, p. 472–487).

Plutonic Rocks

Metamorphosed Dikes and Sills

In scattered areas underlain chiefly by the Orfordville formation scores of metamorphosed dikes and sills, once diabases, have been converted to amphibolites or chlorite schists. Such areas occur a mile north of the village of Norwich, Vt., 2 miles north of the village of Plainfield, N. H., and 4 miles northeast of that village (Pl. 1). Although they have been metamorphosed to a grade identical with that of the enclosing wallrock, relict chill structures and ophitic textures attest to the original nature of the rocks. The maximum mapped length of any single dike is half a mile, and the maximum width 25 feet.

Some of these intrusives have possibly served as feeders for the Post Pond volcanics; in other parts of New Hampshire and Vermont, however, similar metamorphosed intrusives occur in rocks of all ages up through the Devonian Littleton formation (Billings and White, 1950, p. 662). Metamorphosed intrusives of several ages are probably represented in the Hanover area, although there is no way to date them.

Mineralogical assemblages within the dikes represent equilibrium, and examples of the amphibolite, albite-epidote-amphibolite, and greenschist facies are present. A calculation of the chemical composition of any of these dikes, based on the modal analysis, closely approximates that of a diabase. Calcite, ankerite, quartz, and tourmaline are introduced minerals in some of the intrusives, but their relative rarity indicates an inconsequential change in bulk chemical composition of the rocks during regional metamorphism. The generally low quartz content (0–5%) contrasts strongly with that of the Post Pond volcanics (0–20%).

Hornblende Gneiss

Cropping out two-thirds of a mile north of the village of Hartland, and extending northward for 2 miles from this point, is a 200-foot thick sill of hornblende gneiss. Because of its size, its unique mineralogy, and the fact that its relations to the other plutonic rocks are indeterminate, this has been indicated separately on the geologic map.

The rock consists essentially of hornblende, sodic oligoclase, and biotite, with minor quartz, garnet, accessories, and 5 per cent of coarse blue tourmaline. The original character of the intrusive is uncertain, but it probably represents a metamorphosed and tourmalinized diorite or lamprophyre.

Metasedimentary (?) Dike

In an area now submerged under a dam half a mile east of Wilder, Vt., an irregular dike truncates schists of the Orfordville formation. In addition to 5 per cent pyrite the rock consists of approximately equal proportions of dolomite (ω = 1.690), quartz, and muscovite. Its mineralogical similarity to the Waits River formation is striking, and the location of the dike close to the footwall of the Ammonoosuc thrust suggests that it is possibly a part of that formation plastically injected during the faulting. Fowler-Billings (1944, p. 1255–1278) has described actinolite-biotite granulites of the Mt. Washington quadrangle which she has likewise interpreted as metasedimentary dikes.

Diabase

Mildly altered diabase dikes consisting of labradorite, titaniferous augite, some hypersthene, and accessory sphene, apatite, and magnetite are found at several localities within the Hanover quadrangle. The distinctive nature of the augite suggests alkalic affinities, and the dikes are probably representatives of the Mississippian (?) White Mountain magma series. The degree of alteration of the dikes is variable, but almost all of them contain appreciable quantities of calcite, sericite, chlorite, pyrite, and quartz—the same minerals that signal retrogression in the metamorphic rocks. Whether this indicates that regional retrogression in the schists is Mississippian or younger is a matter of conjecture. Because of their relative rarity, none of the diabases is indicated on Plate 1.

Ultramafic Rocks

The tectonic significance of ultramafic rocks makes the question of their location important. The bodies reported by C. H. Hitchcock (1861, p. 539, 543) in the Hanover quadrangle have proved, on examination, to be misidentified. One ultramafic, however, has been located at the northeastern edge of the Lebanon pluton, in the Mascoma quadrangle, 1 mile northwest of Hanover Center. Here a small podlike intrusive less than 10 feet square in outcrop area injects the schists of the Orfordville formation. The rock now consists of 15 per cent anthophyllite ($\alpha = 1.615$; $\beta = 1.630$; $\gamma = 1.640$), 15 per cent talc, 5 per cent magnetite, 5 per cent ankerite, and 60 per cent chlorite (penninite). The ultramafic was apparently metamorphosed concurrently with the enclosing schists.

Four small bodies of amphibolite are intrusive into the rocks on the western contact of the Lebanon pluton (Pl. 3). On the basis of their compositions, three of these bodies are metamorphosed gabbro, but one is a metamorphosed mica pyroxenite. These rocks are not true ultramafics and are possibly comagmatic with the Lebanon granite.

Lebanon Granite

GENERAL STATEMENT: The Lebanon granite (Pl. 3) is an elliptical pluton with a maximum length, in a northeasterly direction, of 6 miles and a maximum width of 3 miles. It was first studied by Merritt (1913) who pointed out the domal nature of the intrusive and the fact that it consisted of two major units—a central core of weakly foliated pink microcline granite and shell of gray granodioritic and quartz dioritic "border gneiss" (Merritt, 1913, Pl. 2).

C. A. Chapman (1939, Pl. 1) remapped the eastern half of the pluton in his study of the Mascoma quadrangle but did not study it intensively. In 1938, Kaiser demonstrated that a considerable amount of granitization had occurred within and adjacent to the pluton. He published a structural map of the body, a series of modal analyses, and petrofabric diagrams of rocks within and outside the granite. The writer agrees with most of Kaiser's conclusions, except for the timing of the intrusion and the origin of the structures in the granite. Kaiser believed that the granite was pretectonic and that its internal structures were superimposed during the deformational episode; the writer is of the opinion that the understanding of the major and minor areal structures hinges upon the fact that the granite is synkinematic.

PETROGRAPHY: The pink granitic core of the Lebanon granite consists chiefly of quartz, microcline-perthite, and sodic oligoclase; epidote, biotite, muscovite, chlorite, zircon, apatite, sphene, allanite, and ores are the minor constituents. In several respects the granite appears to be a normal intrusive. Miarolitic cavities are common at Quarry Hill in Lebanon; a small, poorly exposed offshoot of the granite cuts schists of the Post Pond member of the Orfordville south of the village of West Lebanon (Pl. 1); aplites are common throughout the pluton. On the other hand the granitic core grades into the border gneiss, and this, in turn, into the surrounding metamorphics. Microstructures are complex. Quartz is not uniformly scattered throughout the groundmass but occurs in coarse podlike aggregates. Microcline phenocrysts are commonly replaced by plagioclase, with an accompanying development of myrmekite, but the replacement of plagioclase by microcline is equally common, and complex "perthitic" feldspar intergrowths are the rule in the rock as a whole. Epidote is relatively abundant and crystallizes as discrete grains clearly unrelated to the plagioclase. Its

presence as a primary mineral in the granite doubtless indicates a low temperature of crystallization for that rock. Ramberg (1949, p. 31) has concluded that the maximum stability temperature of epidote is of the order of 400°–600° C.

An increase in the amounts of plagioclase, biotite, epidote, and hornblende, and a decrease in the amount of microcline distinguishes the border gneiss from the granite core. Kaiser (1938, p. 116–125) has presented detailed evidence proving that most of the border gneiss represents a feldspathized schist.

The contact of the border gneiss with the metasomatized Orfordville gneisses and schists is relatively well marked and is determined in most places by a transition from a gneissic to a schistose texture, a decrease in granularity, and a decrease in the feldspar content of the Orfordville rocks. There is no chill zone at this contact. Near Etna village, border gneiss grades along its strike into the metasomatized Orfordville rocks (Pl. 3; *cf.* Kaiser, 1938, p. 127), and the contact of the two units is indefinite. Moreover, some of the metasomatized rocks at the south end of the pluton are strongly gneissic and, except for the presence within them of recognizable beds of amphibolite and biotite schist could be grouped petrographically and structurally as a part of the border gneiss.

Feldspathization of the Orfordville gneisses and schists is indicated by several criteria, in addition to the gradation of the Lebanon border gneiss into them. Intergranular filling by feldspar of the interstices of rounded quartz grains in blastopsammitic schists, complex plagioclase-microcline intergrowths and replacements, quartz-feldspar microveinlets, and the striking variability in the mineralogical composition of the gneisses all point toward metasomatic addition of feldspar, chiefly plagioclase. The petrographic features which typify the feldspathization of the Orfordville rocks are understandably, quite similar to those which led Kaiser (1938, p. 121–125) to conclude that the Lebanon border gneiss is a 'granitized') rock rather than a true intrusive.

How much of the Lebanon granite is "granitized" granite, and how much is magmatic granite is debatable, in view of the conflicting evidence. The writer is of the opinion that much of the border gneiss of the Lebanon pluton is of metasomatic origin; whether the central pink granite core is at all metasomatic cannot be decisively determined.

TABLE 4.—ESTIMATED MODES OF ROCKS OF LEBANON PLUTON

Rock	1	2
No. of sections	7	8
Quartz	35	28
Plagioclase	21	33
Microcline	35	10
Muscovite	2	2
Biotite	4	14
Chlorite	t	t
Hornblende	..	3
Clinozoisite	2	10
Apatite	1	x
Zircon	x	x
Allanite (?)	t	t
Tourmaline	t	..
Sphene	x	x
Magnetite	x	x
Ilmenite	x	x
Calcite	t	t
% An in plag.	11–12	14–30

x = present in 50% or more of thin sections
t = present in less than 50% of thin sections, or in very minor amounts in some sections

1. Lebanon granite
2. Lebanon border gneiss

Modes of the rocks of the Lebanon pluton are indicated in Table 4.

Gneiss at White River Junction, Vermont

Extending for 2 miles both north and south of White River Junction, Vt., with a maximum width of 1 mile, is a *lit-par-lit* complex of chlorite schist intimately penetrated by a gray to pink gneissic "granite." Individual septa within the complex range from a few inches to over 100 feet in width, and "granitic" rocks underlie approximately two-thirds of the area shown as gneiss in Plate 1. Foliate and linear structures are brought out by the orientations of the micas; these structures are strongly distorted near shear zones and near the Ammonoosuc fault. Metamorphosed diabase dikes

(now composed of actinolite, green biotite, chlorite, clinozoisite, sphene, and albite) cut both the "granite" and chlorite schist, indicating that all three rocks have been metamorphosed. The "granitic" rocks, in this respect are considerably different from the Lebanon granite, in which only unmetamorphosed diabase dikes occur.

Petrographically, the "granite" is properly classed as a sodaclase tonalite, for it consists chiefly of albite and quartz, with minor biotite, zircon, clinozoisite, allanite, and pyrite. No potash feldspar has been found in immersions of the powdered rock despite the peculiar microclinelike twinning of much of the feldspar. Two generations of mineral formation are evident. Early subporphyritic albite has been granulated and sericitized, and then penetrated by quartz-albite-sericite-allanite veins. Some of this late albite grows in parallel orientation with the earlier crystals, but it is commonly graphically intergrown with second-generation quartz.

Contacts between the chlorite schist and the tonalite are commonly marked by a transition zone several inches in width, in which microveinlets of quartz, and minor calcite, sericite, and epidote have been injected into the folia of the chlorite schist to produce a gneiss strikingly similar to the gneiss east of Plainfield village. The quartz is strained, and its introduction evidently pre-dated the metamorphism of the *lit-par-lit* complex. The chlorite schist (all of which show petrographic evidence of the introduction of some quartz and carbonate) is composed of albite, chlorite, clinozoisite, quartz, sericite, calcite, ilmenite, pyrite, and apatite. Like the sodaclase tonalite and the metamorphosed diabases, it is an equilibrium member of the greenschist facies. Estimated modes of the rocks of this complex are listed in Table 5.

The probable history of the White River gneiss *lit-par-lit* complex is as follows: (1) folding of the Post Pond volcanics and the development of axial plane cleavage, (2) forcible injection of a "granitic" magma into the cleavage planes, and the wedging aside of the surrounding rocks, (3) intrusion of diabase, (4) low-temperature regional metamorphism, accompanied by the introduction of quartz, albite, allanite, and sulfides, and (5) some subsequent shearing, particularly near the postmetamorphic Ammonoosuc thrust fault. Sodaclase tonalites are described by Johannsen (1932, vol. 2, p. 373) from only one locality and are difficult to account for by almost any theory of petrogenesis. Gilluly (1933, p. 65–81) who has described several such rocks as "albite granites" from the Baker quadrangle of Oregon, has concluded that they are the result of albitization of normal tonalites, the sodic solutions having been derived from an extremely hydrous rest magma by a process of filter pressing. Such an origin, though possible, cannot be proven for these sodaclase tonalites.

TABLE 5.—ESTIMATED MODES OF GNEISS AT WHITE RIVER JUNCTION, VERMONT

Rock	1	2	3	4	5
No. of sections	7	5	3	1	2
Quartz	38	30	10	49	...
Plagioclase	50	45	45	45	40
Calcite	t	t	2	...	...
Actinolite	...	...	...	...	30
Muscovite	6	4	...	2	...
Biotite	2	3	t	1	5
Chlorite	2	10	40	1	10
Clinozoisite	2	8	3	2	10
Apatite	x	x	x	...	...
Zircon	x	...	...	...	...
Allanite	t	t	...	t	...
Sphene	x	t	...	...	5
Ilmenite	t	t	x	...	...
Pyrite	t	t	x	t	...
% An in plag.	4–8	4–7	4–8	5	6

x = present in 50% or more of thin sections
t = present in less than 50% of thin sections; or present in very minor amounts in some thin sections

1. Sodaclase tonalite
2. "Intermediate" rock
3. Chlorite schist
4. Aplite
5. Metamorphosed diabase, intrusive into the sodaclase tonalite

The age of this gneiss cannot be accurately determined through geologic evidence. Structurally and petrographically, however, it is

similar to rocks of the Highlandcroft magma series (Billings, 1937, p. 500). Its probable age, therefore, is late Ordovician.

Gneiss East of Plainfield, New Hampshire

A gray gneiss whose linear and foliate structures are conformable with those of the surrounding Post Pond volcanics crops out in an area 2 miles northeast of the village of Plainfield. Contacts with the wall rock are gradational over a distance of a few hundreds of feet, and even within the area indicated as gneiss (Pl. 1) there are numerous "inclusions" or beds of chlorite schist. In one small area, near the northeastern contact of the body, there is a sodaclase tonalite identical with that which forms the gneiss complex at White River Junction, but most of the gneiss is a chlorite-albite-clinozoisite-quartz rock in which hornblende is sparingly present. Diabase dikes intruding the Plainfield gneiss have been metamorphosed, and metamorphic isograds cut across the geologic boundaries (Fig. 11); thus the complex is pre-metamorphic.

The outstanding microscopic feature of the Plainfield gneiss is a vein and podlike structure of the quartz, which is considerably coarser than the other minerals of the ground mass. Associated clinozoisite, calcite, and ankerite are present in highly variable amounts in these microveinlets. In those rocks that contain micas, quartz is always present; some quartz-bearing rocks, however, contain no mica. Both megascopic and microscopic details suggest this gneiss was in part formed by a "soaking" of the Post Pond volcanics by siliceous emanations carrying minor carbonate and, probably, potassium. Some of the volcanics may have initially been siliceous, and sedimentary rocks may have been present before the series was transformed, but some of the quartz now present represents, on the basis of its petrographic nature, introduced material. Estimated modal analyses of the rocks described are listed in Table 6.

The estimated average chemical composition of this gneiss may be compared with that of the Post Pond volcanics (Table 7). For numerous evident reasons these analyses are grossly inaccurate. In the calculation of the composition

TABLE 6.—ESTIMATED MODES OF GNEISS EAST OF PLAINFIELD, NEW HAMPSHIRE

Rock	1	2	3
No. of sections	1	10	2
Quartz	40	32	21
Plagioclase	55	40	42
Calcite	....	2	...
Ankerite	....	t	2
Sericite	....	3	...
Biotite	....	2	1
Chlorite	....	13	3
Hornblende	....	..	25
Clinozoisite	3	8	4
Apatite	....	x	x
Zircon	....	t	...
Sphene	....	t	...
Magnetite	....	x	2
Ilmenite	....	t	...
Pyrite	2	t	...
An in plag.	11	6–24	24–27

x = present in 50% or more of thin sections
t = present in less than 50% of thin sections; or present in very minor amounts in some thin sections

1. Sodaclase tonalite
2. Chlorite-albite-clinozoisite gneiss
3. Hornblende schist

TABLE 7.—ESTIMATED AVERAGE CHEMICAL COMPOSITION OF GNEISS EAST OF PLAINFIELD AND OF POST POND VOLCANICS

	Post Pond	*Post Pond (less 20% SiO_2)*	*Daly's Basalt*	*Plainfield gneiss*	*Daly's granodiorite*
SiO_2	60.84	50.84	49.06	67.20	65.01
TiO_2	.30	.38	1.36	....	.57
Al_2O_3	12.18	15.22	15.70	14.70	15.94
Fe_2O_3	2.93	3.66	5.38	0.29	1.74
FeO	5.65	7.06	6.37	1.61	2.65
MgO	4.72	5.90	6.48	2.93	1.98
CaO	6.41	8.01	8.95	5.27	4.42
Na_2O	3.15	3.93	3.11	4.30	3.70
K_2O	0.34	.42	1.52	.90	2.75
H_2O	1.76	2.20	1.62	1.90	1.04
CO_2	.80	1.00	....	.90	.....
P_2O_5	.45	.56	.45	....	.20
	99.53	99.18	100.00	100.00	100.00

of the gneiss east of Plainfield, the modes of Table 6, column 2 were used; for the Post Pond volcanics, columns 1, 2 and 5 of Table 3 were assumed to most closely approximate the average composition of the volcanics.

The Post Pond volcanics have an average composition which is four-fifths basalt and one-fifth quartz, but the Plainfield gneiss roughly approximates a granodiorite.

Structure

General Statement

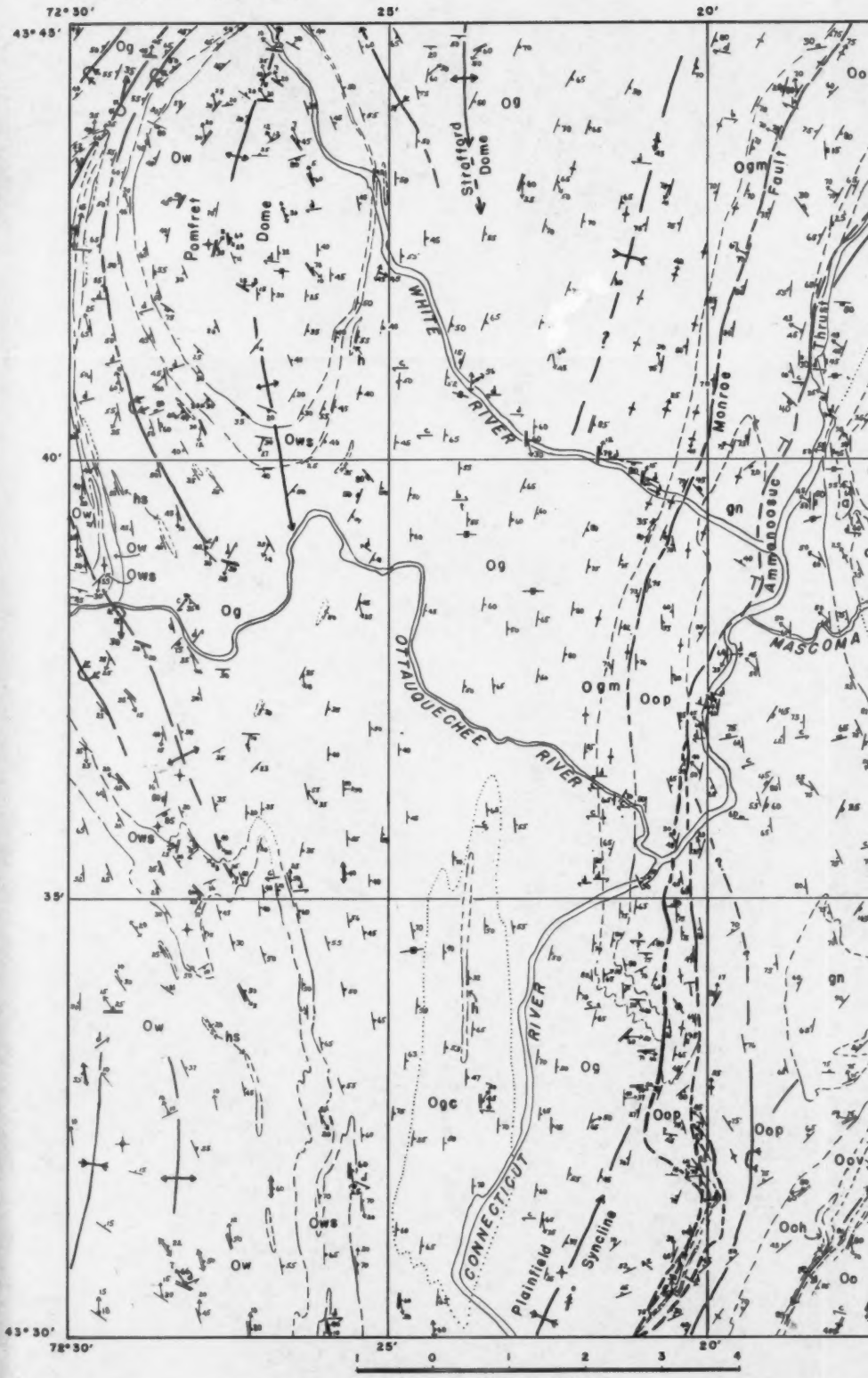
The structural environment of the Hanover quadrangle is illustrated by Figure 4. The eastern limb of the Green Mountain anticlinorium is characterized by a thick homoclinal sequence of Cambro-Ordovician eugeosynclinal metamorphics intruded by serpentines which themselves form a continuous belt from Massachusetts to eastern Quebec. Some 20–30 miles west of Hanover this homoclinal sequence steepens and passes into a tight syncline, east of which is a remarkable series of cleavage domes. Ten to fifteen miles east of the cleavage domes are the granitic domes of the Oliverian magma series. Both the cleavage domes and the Oliverian domes have an en echelon pattern, each individual dome striking almost due north and being offset to the northeast relative to the dome south of it. The tremendous sheetlike Mt. Clough pluton marks the eastern limit of the Oliverian domes; east of this are the widespread stocks, batholiths, and ring dikes of the New Hampshire and White Mountain magma series.

The Hanover area lies between the cleavage domes on the west and the Oliverian domes on the east. The most obvious structure in the rocks, the cleavage, forms a fan in cross section, the rocks in the western part of the area dipping easterly, and those in the eastern part of the area dipping westerly. The axis of this cleavage fan lies slightly west of the Connecticut River throughout its course. Three large anticlines dominate the area, the Pomfret dome, the Lebanon dome, and the Meriden anticline (Pl. 2). Three faults have been traced into the area from the northeast; the Monroe, Ammonoosuc, and the Northey Hill thrusts. Folding in all the rocks is complex, and in the western part of the area recumbent folds have developed. Although a clear-cut separation of the folds into two generations is not possible throughout this area, as it is in near-by portions of eastern Vermont (White and Jahns, 1950, p. 198–200), part of the structural complexity is demonstrably due to a superposition of later minor folds on earlier ones. Detailed structural readings are recorded on Plate 2.

Pomfret Dome

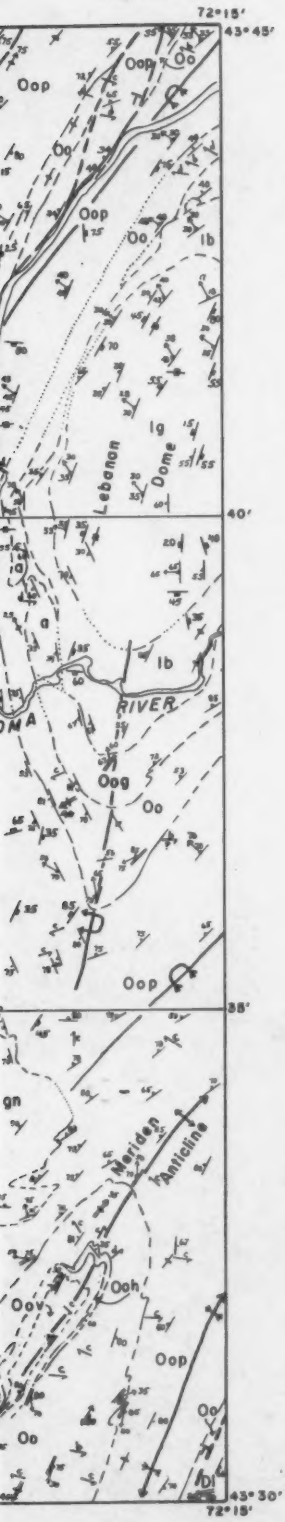
A typical representative of the cleavage domes of eastern Vermont is outlined by the map pattern of the Standing Pond amphibolite in the northwestern ninth of this quadrangle. Like all other cleavage domes this has a deceptively simple map pattern; the inner structure of the dome is actually complex, and, instead of bedding dipping quaquaversally outward, the axial-plane cleavage of a series of isoclinal folds dips outward and conforms to a domal pattern. In the Pomfret dome bedding and cleavage in the Standing Pond amphibolite and the basal Gile Mountain formation are parallel and dip quaquaversally outward. The Waits River formation in the interior of the dome, however, has been crumpled into a series of minor folds whose axial-plane cleavages parallel the bedding cleavages in the other two formations. Thus the folding is highly disharmonic. Exposures of recumbent folds are not sufficiently good or plentiful to allow a consistent deduction of their movement sense, but in one glacially plucked 100-foot cliff 1.5 miles N. 55° W. of West Hartford village, a bed of calcareous schist has been repeated by folding five times, and there is no obvious drag pattern.

Geological mapping indicates that the Pomfret dome is symmetrical, but Bean (1951, p. 85), from a study of gravity anomalies, has detected a steep-walled mass deficiency beneath the eastern side of the structure. Its depth is unknown, owing to uncertainty about the detailed stratigraphy in this portion of Vermont. The Strafford dome to the northeast has, however, a mass deficiency at approximately 2650 feet below the surface; it is probable (Bean, 1951, p. 85) that the depth to the mass deficiency of the Pomfret dome is of a similar order of magnitude. That the mass



TECTONIC MAP OF THE HANOVER QUADRANGL
NEW HAMPSHIRE-VERMONT

Lyons, Pl. 2



LEGEND

STRUCTURAL SYMBOLS

- 10/ Strike and dip of beds
- 10 Strike and dip of overturned beds
- 20 Strike and dip of foliation or schistosity
- Strike of vertical foliation or schistosity
- Horizontal schistosity
- 30 Strike and plunge of linear element
- Strike of horizontal lineation
- 55 Strike and dip of axial plane of minor fold
- 20 / 30 Strike and dip of axial plane of minor fold, with strike and plunge of fold axis
- 50 Strike and dip of axial plane of minor fold, with horizontal fold axis
- b Top of formation as deduced from
  b - primary bedding features
  c - cleavage - bedding relations
  d - drag folds
  p - pillow structure
- Anticline, showing trace of axial plane and bearing and plunge of axis
- Syncline
- 30 Overturned anticline, showing trace of axial plane, direction of dip of limbs, and bearing and plunge of axis
- Overturned syncline
- 30 Strike and dip of joints
- Fault

Formation symbols as on geologic map (Plate I)

NGLE.

deficiency is probably granite is indicated by: (1) an increase in the intensity of metamorphism toward both the Pomfret and Strafford domes, (2) a small stock of granite 4 miles north of the Strafford dome, and along its projected axis (Doll, 1944, Pl. 1), and (3) a small pegmatite dike cutting through the Pomfret dome at the locality 1.5 miles N. 55° W. of West Hartford village. The significance of this dike is brought out by the fact that pegmatite dikes are otherwise unknown in the Vermont portion of the Hanover quadrangle.

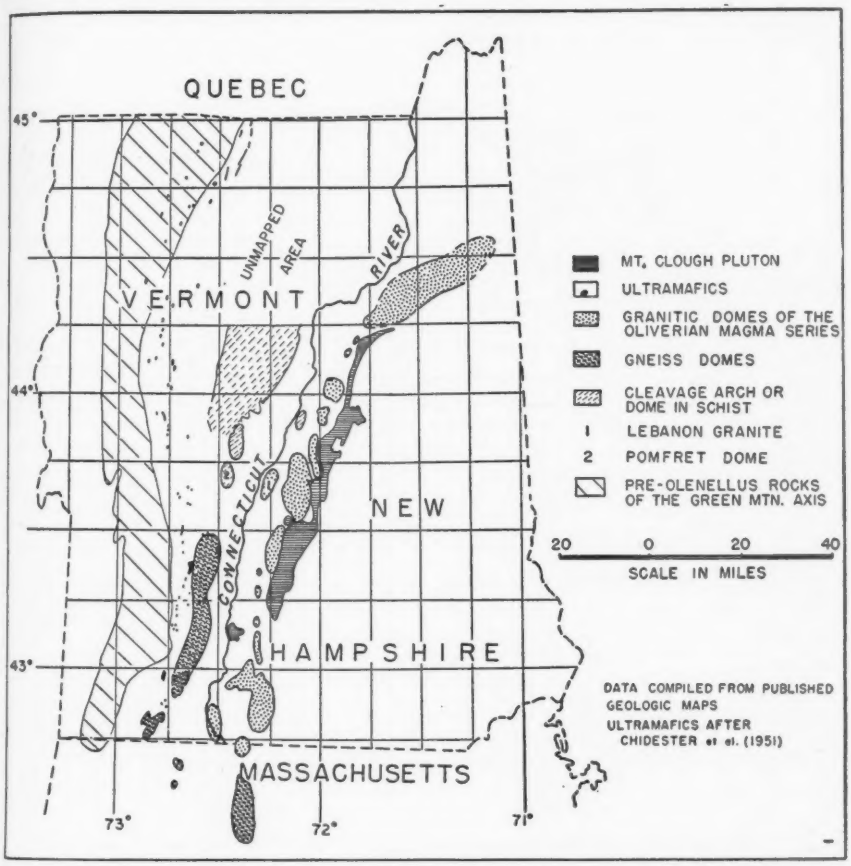


FIGURE 4.—SKETCH MAP OF DOMAL STRUCTURES OF NORTHWESTERN NEW ENGLAND

It has been demonstrated by White (1949, p. 590) and White and Jahns (1950, p. 210) that the cleavage in both the Strafford dome and the eastern Vermont cleavage arch (which lies north of it) can be correlated with a late-stage deformation which followed a period of north-northeasterly folding. This later deformation is recognizable because, to the north, it is marked by a northwesterly-trending slip cleavage which cuts across the earlier structural trend and merges into the cleavage of the cleavage arch. Clear-cut evidence of such second-deformation structure is not apparent around the Pomfret dome, possibly because the angle of strike between the earlier schistosity and the later slip cleavage is so acute that evidence of second-stage slippage is not detectable. It is nevertheless probable, in the light of

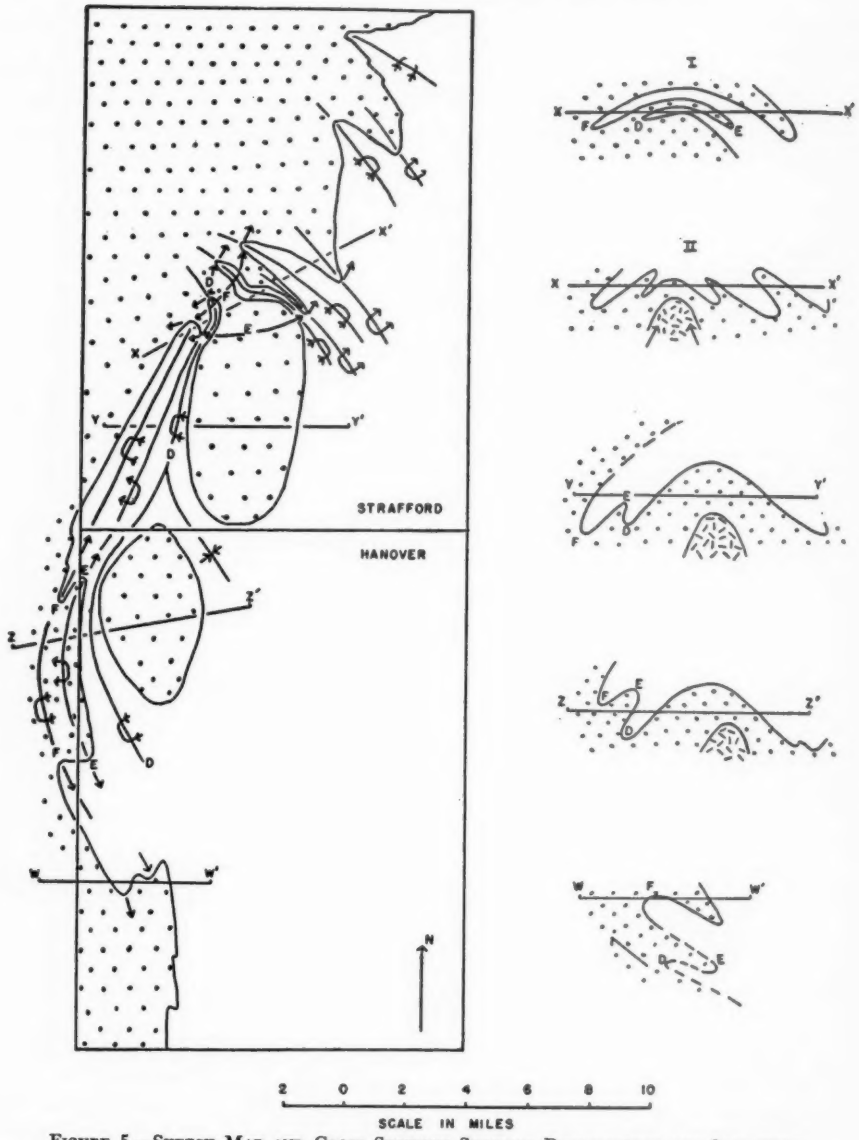


FIGURE 5.—SKETCH MAP AND CROSS SECTIONS SHOWING DISTRIBUTION AND STRUCTURE OF STANDING POND AMPHIBOLITE

Dark line at the upper contact of the Waits River marks the approximate position of the Standing Pond amphibolite.

Stippled pattern = Waits River formation
White area = Gile Mountain formation
Dashes = Igneous rocks (inferred from gravity data)
Arrows = Direction of plunge of minor folds

I. Section along line X–X′, modified from White and Jahns (1950, Fig. 11). Letters correspond to their labelled fold axes at north end of Strafford dome. This section approximately parallel to fold axes.

II. Alternative interpretation (after Doll, 1944) of structure at north end of Strafford dome. This section disregards direction of plunge of minor fold axes.

White and Jahns' data, that the formation of the Pomfret cleavage dome post-dated the initial regional deformation.

Figure 5 shows the overturning of the folds toward the Strafford and Pomfret domes, and some significant differences in the interpretation of the structure, particularly at the north end of the Strafford dome. Doll's (1944, p. 23–26) idea of a simple cross-folded anticline is shown in section II, and by contrast section I is the complicated nappe interpretation of White and Jahns (1950, p. 213) based upon the assumption that the direction of plunge of the minor folds is of major structural significance.

In the White-Jahns (1950, p. 213) concept of the north end of the Strafford dome, folds D and F of Figure 5 are the noses of two nappes overfolded from the east to the west. The axes of folds D, E, and F are traceable almost into the southwestern ninth of the Hanover quadrangle where these folds are replaced by a series of folds overturned toward the Ludlow dome on the southwest (Fig. 6a and 6b). Whether nappelike overfolds are present west of the Pomfret dome is not demonstrable on the basis of the available field data. Probably, however, the nappe structure is nonpersistent and dies out on the Strafford dome itself.

Before considering the origin of the Pomfret dome, certain features common to all the cleavage domes should be briefly cited. All possess a disharmonic fold pattern which is probably due to extreme flowage of the less competent rocks. A gravity low (Bean, 1951, Pl. 1) is present over each dome that has been investigated geophysically. Minor folds on the flanks of the domes have, in many cases, a "reverse" pattern evidently due to an upward welling of the central part of the dome. This upward movement has been so intense in the Chester dome (Thompson and Rosenfeld, verbal communication) that a major syncline west of that structure has been involved and dragged into an inverted position. Slip cleavage grades into the schistosity of the domes north of the Hanover quadrangle; there, too, the slip cleavage indicates relative upward movement of the central part of the dome.

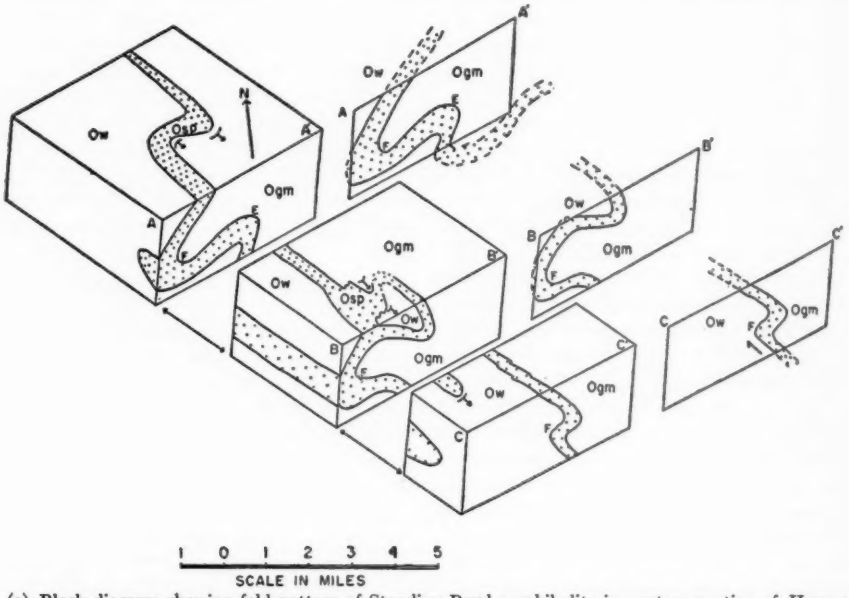
White and Jahns (1950, p. 214–219) have concluded that squeezing of an initially folded group of schists has resulted in the upward flowage of the calcareous rocks and the development of the cleavage domes or arches. Their mechanism, which is difficult to visualize, is rendered improbable by Bean's (1951, p. 85) discovery of low-density rocks in the center of the domal structures. A more likely explanation, and one more consistent with current data on all the domes, would be that subsequent to the initial folding of the rocks, or perhaps in the late stages of folding, low-density rock (granite or migmatite) started to rise. The ascent of this rock became the dominant late-stage tectonic event and accounts for most of the structural complexities associated with the domes.

Recumbent Folds of Western Part of Quadrangle

The Standing Pond amphibolite, traced southward from the area west of the Pomfret dome, undergoes an intricate series of convolutions which are best understood by reference to the tectonic map (Pl. 2) and to Figure 6. As is shown by section B-B′ of figure 6a the amphibolite has locally been overturned through an angle of more than 180° and rests, in a nappelike fold 1½ miles across, upon the inverted Gile Mountain formation. This complication occurs where there is a sudden transition from the easterly overturning on the western side of the Pomfret dome to the westerly overturning on the northeastern side of the Chester dome. As has been discussed previously, the latter dome shows evidence of strong upward movement of the central part of the structure. It is believed that the explanation of the complicated relations shown by figure 6a is that syncline F of section B-B′ has been dragged upward and westward so that it now occupies the nose of a reverse drag fold (section C-C′, Fig. 6a). The inversion, by dragging, of fold F explains the nappelike overfold pictured in the drawing of figure 6b. This conclusion, the only one consistent with the carefully mapped structures, is in agreement with the statement of White and Jahns (1950, p. 216) that somewhat similar folding at the north end of the Strafford dome "points to a common upward movement of rocks of the central tectonic belt" (*i.e.*, the cleavage arch of eastern Vermont and the Strafford dome). It is important to note that the explanation of the geology of figure 6a and 6b involves two separate orogenic episodes. The first folded the rocks into north-

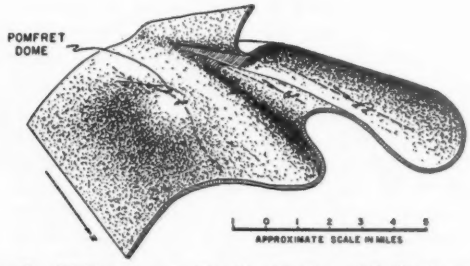
south anticlines and synclines; the second is related to the rise of the cleavage domes and introduces the structural complexities. As has been pointed out by White and Jahns (1950, p. 214–219) there is no way of estimating the time lapse between the deformational episodes. Conceivably they may be part of the same major orogeny.

Despite the complexity of the nappelike overfold and the evidence of upward drag toward the west, the Waits River formation in the southwestern ninth of the Hanover quadrangle appears to be warped into a series of relatively gentle, poorly defined anticlines and synclines, overturned toward the west. That the true structure is far more involved than is deducible from scattered outcrops is apparent from cliff outcrops which expose small recumbent folds. These folds have themselves been warped into the gentle structures now shown on the tectonic map.



(a) Block diagram showing fold pattern of Standing Pond amphibolite in western portion of Hanover quadrangle.

Og = Gile Mountain formation
Ows = Standing Pond amphibolite
Ow = Waits River formation



(b) Sketch, looking south, of fold pattern in Standing Pond amphibolite in western portion of Hanover quadrangle. Drawing by J. L. Snyder.

FIGURE 6.—RECUMBENT FOLDS

Meriden Anticline

The doubly plunging anticline which underlies most of the southeastern ninth of this quadrangle is a relatively simple structure, whose axial plane dips steeply westward. Two unusual features are noteworthy: (1) the eastern limb has a superimposed set of steeply plunging minor folds which are not parallel with the major fold axis, and which have an east-side-north movement-sense; (2) as the fold axis is projected into the Mascoma quadrangle, the Hardy Hill quartzite reappears along it. Reasons for these relations will be advanced under the discussion of the Lebanon dome.

Lebanon Dome

The oval area underlain by the Lebanon granite and the Lebanon border gneiss has been described as a dome by Merritt (1913, p. 8) and classed as a member of the Oliverian magma series by Kaiser (1938, p. 109) and Chapman (1939, p. 143). It is referred to in this paper as a dome for convenience of discussion, although the pluton actually has the shape of a cone or cylinder tilted toward the northwest (Pl. 3).

Foliate structures in the granitic rocks are brought out by the parallel alignment of micas, of schlieren, and, in some places, of tabular feldspars. Lineation in the foliation planes is due to a streaking-out of micaceous clots and elongation of the schlieren. Both types of internal structure parallel foliate and linear structures in the surrounding Orfordville formation and in the four amphibolite intrusives along the western contact of the Lebanon pluton.

The granite core of the pluton is offset to the west of what, from its map pattern, should be be the axial trace of the "dome." Moreover, the foliate structures in the granite itself dip westerly, so that the eastern limb of the pluton is overturned or very nearly vertical. Unusual structural relations between the Lebanon border gneiss and the Lebanon granite are likewise evident near the north end of the "dome." In the map of figure 7, which represents a small area 2 miles N. 50° W. of the village of Etna (Pl. 3), the gradational petrographic contact between the core granite and the border gneiss has a southerly trend, but the foliate and linear structures, which here are wrapping around the northern end of the "dome," transect the lithologic boundaries. These relations are explicable in a variety of ways but are probably due either to "granitizing" encroachment of the granite core on the granodioritic border gneiss subsequent to the development of the planar and linear structures in the latter, or to the fact that these structures were not developed until both rocks were semisolid.

Foliation in the Orfordville formation surrounding the granite is either bedding-cleavage or axial-plane schistosity, but lineation develops in a variety of ways. In mica schists or gneisses it is a crinkling structure; in quartz-mica schists and the metasomatized gneisses it is a streaking-out of the micas; and in the hornblende schists of the Orfordville formation and the intrusive amphibolities it is a parallel growth of the hornblende needles. Around the south end of the intrusive, lineation in the schists appears in a series of minor folds with steep northerly plunging axes.

Joints, filled with granitic aplites or hydrothermal minerals, may be grouped, in accordance with the standard nomenclature of granite tectonics, as cross, longitudinal, or diagonal joints. Of these three types the cross joints are by far the commonest and persist outward from the pluton into the country rock.

Structures plotted on Plate 3 indicate that the Lebanon pluton is asymmetric, with a moderately dipping western flank and an almost vertical or overturned eastern contact. The northern end of the body plunges gently north-northwest, but the southern end plunges steeply toward the north, parallel to the plunges of the folds in the Orfordville formation. In other words the entire pluton plunges north-northwesterly and has the shape of a distorted cone. Moreover, the northerly plunging Orfordville formation south of the pluton has been inverted. This interpretation, based upon mappable structures, has been recently corroborated by Bean (1951, p. 91–93) who has made a detailed gravimetric study of the dome. His three cross sections, based upon Bouguer anomalies and drawn to a depth of 12,000 feet below the surface, agree very closely with similarly oriented sections drawn from the structural data.

The fact that the foliate structures of the Lebanon granite and border gneiss outline the shape of the pluton, that the linear structure points in the direction of elongation of the body, the feldspars within the plutonic rocks, the superposition of a foliation on some of the aplites, and the parallelism of the structures within and outside the body means that some

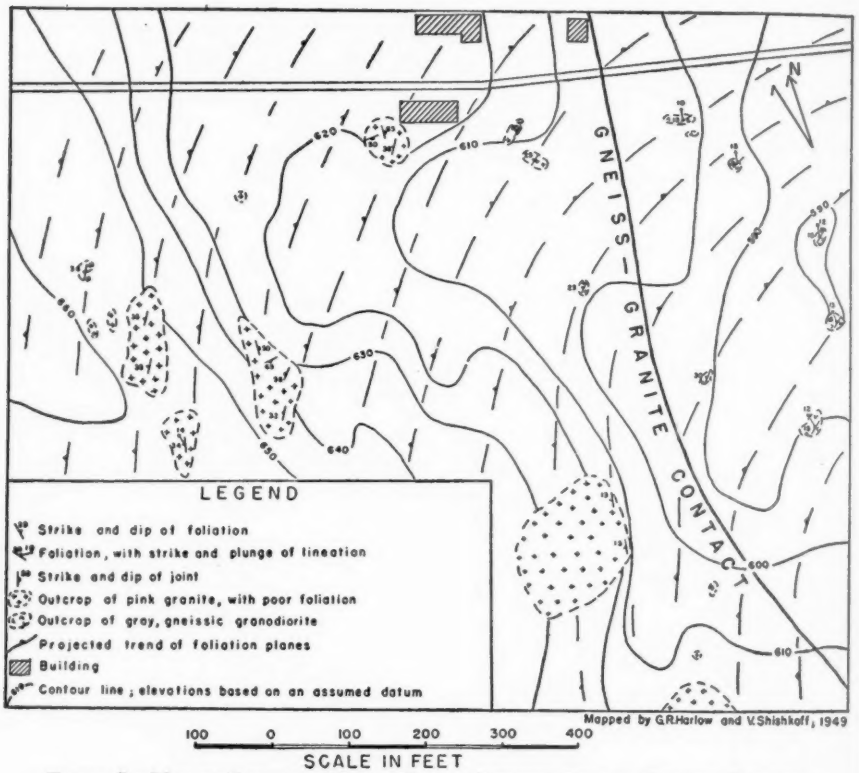
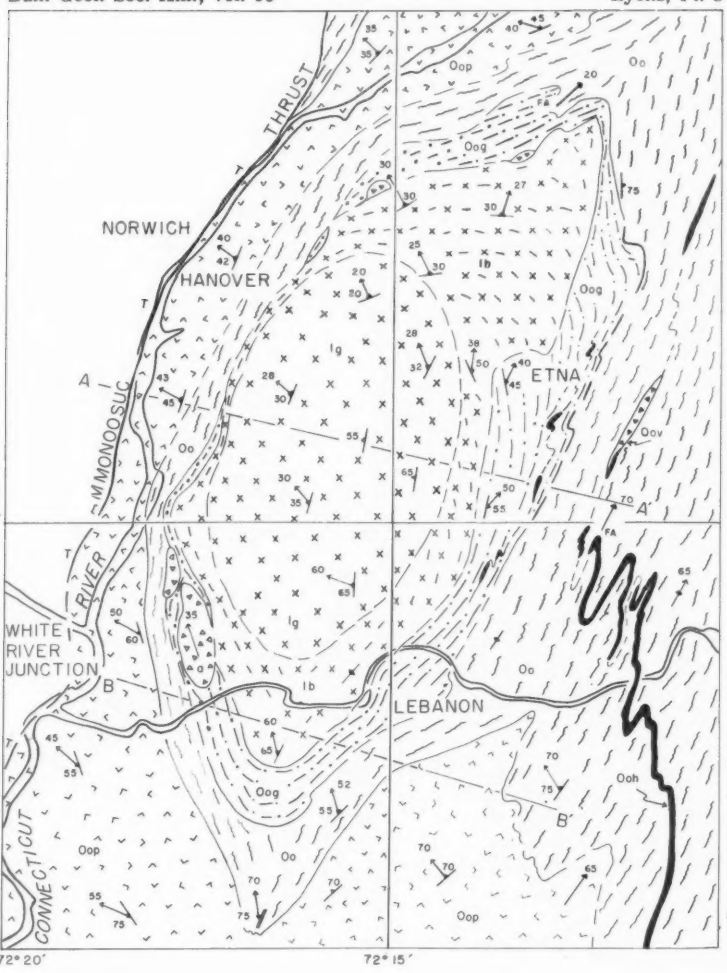


FIGURE 7.—MAP OF PORTION OF NORTH END OF LEBANON PLUTON; BERWICK PASTURE, HANOVER, N. H.

that the schlieren are parallel to the foliation and lineation, and that the joints have orientations predicted by granite tectonics indicate that the lineation and foliation within the granite are primary flow structures. The inversion of the beds south of the body is a reflection of an invasive movement in that direction and is incompatible with any theory which ascribes a secondary origin to foliated granitic structures (*cf.* Kaiser, 1938, p. 126) or which explains their present distribution by the folding of a horizontally foliated granitic sheet (Moore, 1949, p. 1666–1667).

The slight protoclastic structure of some of the rocks moved while solid or semisolid. Recrystallization and metamorphism of the Orfordville formation was evidently synchronous with the emplacement of the granite, as is indicated not only by the concentric pattern of the metamorphic isograds about the Lebanon "dome" (Fig. 11) but also by the parallelism of hornblende needles outside the pluton with linear structures inside it. These crystals are not fractured and show no evidence of having been dragged or rotated into their present position. They must, therefore, have grown in an orientation related to and controlled by the moving granite.



LEGEND

- Ig — LEBANON GRANITE
- Ib — LEBANON BORDER GNEISS
- a — INTRUSIVE AMPHIBOLITE
- Oog — ORFORDVILLE GNEISS
- Oop — POST POND VOLCANICS (ORFORDVILLE)
- Oo — ORFORDVILLE SCHIST
- Oov — ORFORDVILLE VOLCANICS
- Och — HARDY HILL QUARTZITE (ORFORDVILLE)

DEVONIAN(?) — ORDOVICIAN(?)

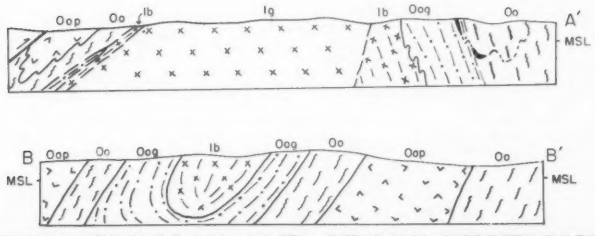
SYMBOLS

- ACCURATE CONTACT
- APPROXIMATE CONTACT
- THRUST FAULT T = UPPER PLATE
- 20 STRIKE AND DIP OF FOLIATION
- 60 STRIKE AND PLUNGE OF LINEATION
- FA 70 STRIKE AND PLUNGE OF FOLD AXES
- 20 STRIKE AND DIP OF AXIAL PLANE OF MINOR FOLD

1 0 1 2 3 4 5

SCALE IN MILES

N



Geology mapped by J.B. Lyons 1947-1949

Easternmost belt of Ooh is after C.A. Chapman (1939, Pl. I)

GEOLOGIC MAP AND CROSS SECTIONS OF THE LEBANON GRANITE

The sy
is brough
tions whi
a conside

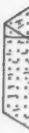


Figure 8
the regi
full len
diagram
third of
ern thi
blocks h
parallel
(Pl. 1).
Hill qua
is the a
and Ma
thrown
plungin
minor f
tions su
ite are

The syntectonic nature of the Lebanon pluton is brought out not only by the structural relations which have already been described but by a consideration of the features indicated in stratigraphic relations and excellent examples of graded bedding and channeling demonstrate that beds toward the southwest are the upper stratigraphic units, and that these minor folds,

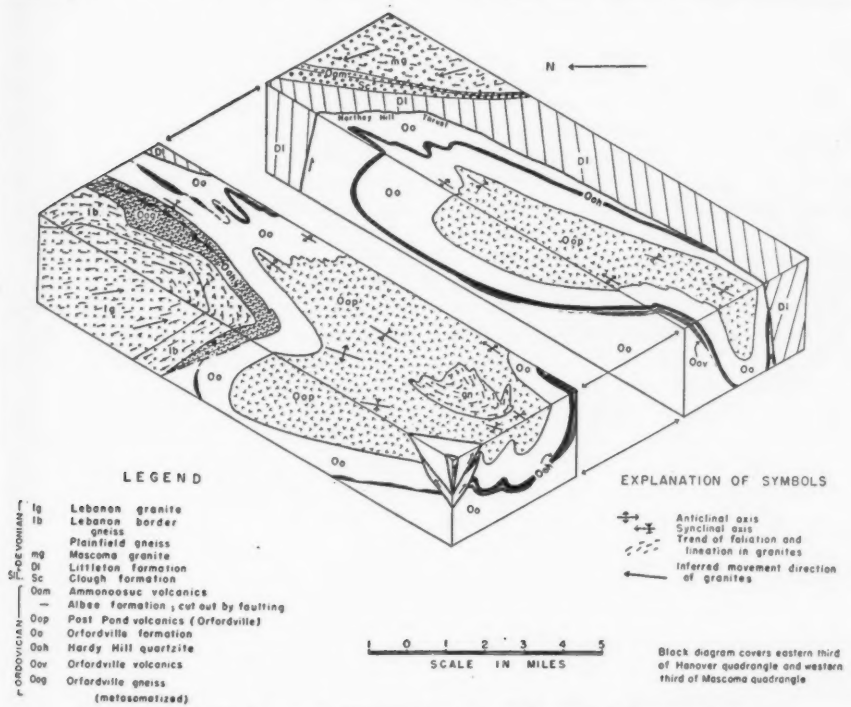


FIGURE 8.—BLOCK DIAGRAM SHOWING AREAL AND STRUCTURE RELATIONS OF LEBANON GRANITE

Figure 8. This block diagram, drawn parallel to the regional structural trend, extends for the full length of a quadrangle and illustrates diagrammatically the geology of the eastern third of the Hanover quadrangle and the western third of the Mascoma quadrangle. The blocks have been pulled apart along a line which parallels the axis of the Meriden anticline (Pl. 1). The key structural unit is the Hardy Hill quartzite, and the most significant locality is the area between the outcrops of the Lebanon and Mascoma granites. Here the quartzite is thrown into a complex series of steep northerly plunging folds. (*Cf.* Chapman, 1939, Pl. 1.) The minor fold pattern and cleavage-bedding relations suggest that beds northeast of the quartzite are stratigraphically younger. Regional like those south of the Lebanon pluton, are inverted. These complex structural relations are most easily explained by assuming that the Lebanon pluton moved upward toward the southeast at the same time that the Mascoma pluton (as indicated by its foliate and linear structures) moved upward toward the west or northwest. If these plutons rose synchronously, rocks between them would necessarily have been involved in a gigantic couple with an east-side-north, west-side-south movement sense. This couple appears to provide the only logical explanation for the inversions of the folds and the peculiar structural relations indicated in figure 8. It likewise explains why, in this region, minor fold patterns and cleavage-bedding rela-

tions are treacherous criteria for the recognition of tops and bottoms of beds.

Whether the rising plutons were the cause of the relative movement indicated by the mapped geologic relations is conjectural. Over large parts of the Hanover, Mt. Cube, and Claremont quadrangles there is a series of minor, steeply plunging folds which have been superimposed on the major folds, and which have an east-side-north drag sense. This suggests that during the emplacement of the Lebanon granite a gigantic horizontal couple of unknown origin was operative and that the granites rose into areas of relatively low stress. Density differences between the plutonic and the country rock and an upward kneeding of the lighter rocks subjected to the stronger stresses at depth provided the mechanism for the forceful injection. If the minor fold pattern, the emplacement of the granite, and the inferred horizontal couple are related in time, as they appear to be, the Lebanon "dome" would have arisen subsequent to the initial north-south folding of the geosynclinal rocks, at a time of horizontal subcrustal adjustment to the deforming stresses.

Structure of Area Between Lebanon and Pomfret Domes

Disregarding the interruptions caused by faulting, the cleavage, as noted previously, forms a fan in cross-section. The axis of this fan lies 1½–2 miles west of the Connecticut River, about midway between the Lebanon and Pomfret domes. In the northern third of the quadrangle a synclinal axis seemingly coincides with the cleavage axis (Pl. 2), but whether a major syncline continues through the entire quadrangle is problematical. This is a matter of considerable importance in the stratigraphy of the area but one which is not resolvable because of the lack of reliable top-and-bottom indicators. The major structural argument against the north-south continuation of the syncline is the geology of the south-central ninth of the Hanover quadrangle, where the excellently exposed Plainfield syncline plunges northeastward (Pl. 2). The position and trend of this fold seem to rule out the possibility of a major northward-trending syncline immediately west of it.

An interesting but puzzling series of outcrops appears about 2 miles south of the mouth of Kilburn Brook, at the western side of the east-central ninth of the quadrangle (Pl. 1) and along both banks of the Connecticut River. Calcareous amphibolite and marble beds on the east bank of the river appear to be gently warped, and sub-horizontal; on the west bank of the river, however, recumbent folds are evident, some of them with as much as 30 feet of both upper and lower limbs exposed. Axes of these folds parallel the major trend of the rocks, but whether the folds open toward the east or the west is not determinable. The interpretation adopted on the geologic map is that these recumbent structures lie on or near the axis of a major syncline (the Salmon Hole Brook syncline?) which lies immediately west of the Lebanon dome. The synclinal axis is truncated by the Ammonoosuc thrust north of the above-described locality, but a few miles to the south it emerges from the thrust and is traceable into the Claremont quadrangle to the south.

Minor Folds

Structural relations throughout this area are complicated by the superposition of numerous minor folds on the major structures. These minor folds, which vary from microfolds to flexures several hundred feet in amplitude, in general parallel the major folds. Commonly, however, the plunge of folds in incompetent beds is as much as 30° steeper than that of the major structure.

Axes of minor folds peripheral to the Lebanon dome and the Meriden anticline plunge concordantly with the major structures, and most of the minor folds can be interpreted as normal drag folds. Exceptions are the previously mentioned steeply plunging minor folds (chiefly present in the eastern side of the area) which are due to an east-side-north slippage along cleavage or bedding planes. Their development is considered a reflection of the horizontal couple produced by or related to the emplacement of the Lebanon granite. In the sense that they have been superimposed on preexisting folds, they represent a second-deformation type of structure.

Minor folds of the western third of the quadrangle show a similar close correspondence between the plunge of the major structures, such as the Pomfret dome, and the plunge of the minor folds. These major structures and many of the minor folds are superimposed on a pre-existing set of anticlines and synclines. The resultant complexity of structure, when coupled with the varying behavior of different types of rocks, makes any reliable resolution of top-and-bottom relations impossible, either on the basis of the minor folds or of the cleavage which ordinarily parallels their axial planes. The highly plastic behavior of the Waits River formation, for example, allows one to deduce only that the beds have been isoclinally folded, for the attenuation and dragging out of the limbs is so extreme that no movement-sense can be inferred from their pattern. The Standing Pond amphibolite and basal Gile Mountain formation are ordinarily so competent that minor folds are not developed; those few present on the western limb of the Pomfret dome have a "normal" pattern which may be related to the initial folding.

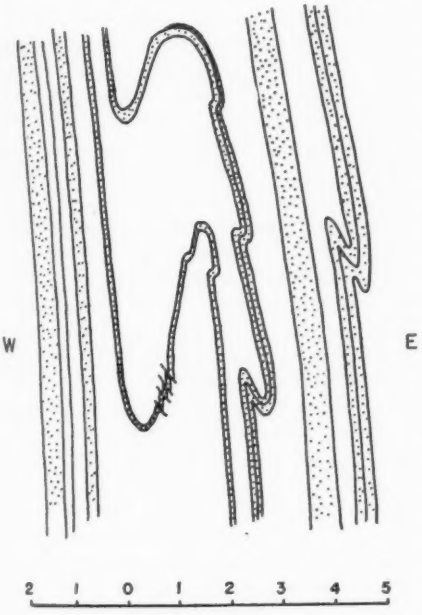


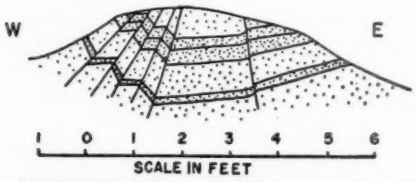
FIGURE 9.—SKETCH OF FOLDS IN THE GILE MOUNTAIN FORMATION
West of village of Hartford, Vt., showing larger normal drag folds and smaller superimposed reverse drags.

Within the Gile Mountain formation of the central part of the Hanover quadrangle numerous minor folds reflect distortions in the major structure. Of particular interest are the folds in the Meetinghouse slate and the easternmost part of the Gile Mountain formation, most of which have a tops-west drag pattern incompatible with the assumption that the Meetinghouse slate is the uppermost member of the Gile Mountain. A possible explanation of the anomaly is indicated in figure 9, a sketch made from a photograph of some excellently exposed folds half a mile west of the village of Hartford, Vt., and along the northern bank of the White River.

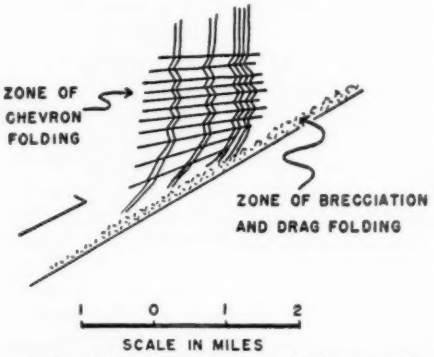
The larger of the two folds is a drag fold on the western limb of a syncline. The syncline itself is part of a group of vertical isoclinal folds (not shown in the sketch) which have wave lengths of 30 feet or more, and no consistent drag sense. The smaller fold in the sketch has a tops-west drag pattern and is most easily interpreted as a second-deformation fold superimposed on a first-deformation fold. It is identical in most respects to the second-deformation minor folds described by White and Jahns (1950, p. 196–200), Hadley (1950, p. 28–32), and White and Billings (1951 p. 681) from areas to the north and is, in all probability, genetically related to the formation of the Pomfret cleavage dome. Other such examples of second-deformation folds are recognizable throughout the Gile Mountain formation, but their identification is difficult. In areas to the north, slip-cleavage parallels the axial planes of these structures and is of a different orientation than the earlier axial-plane cleavage. In the Hanover area this situation is evidently not so, and there are, consequently, no reliable criteria to apply to a single fold to differentiate it as a first- or second-deformation structure. This means that the drag pattern of folds cannot be relied upon to determine top-and-bottom relations, particularly as one proceeds westward across the quadrangle.

A prominent group of second-deformation

folds with wavelengths of up to 20 feet is developed in the Meetinghouse slate and Gile Mountain formation 1½–2 miles north of the village of Plainfield, in the south-central ninth



(a) Chevron folds in Gile Mountain formation.



(b) Sketch showing, diagrammatically, a possible mechanism for development of chevron folds near faults.

FIGURE 10.—CHEVRON FOLDS

of the quadrangle. Here an essentially horizontal couple has thrown the rocks, which once struck north and dipped easterly, into a series of minor folds plunging easterly at approximately 50°. That this folding is a superimposed structure is indicated by the observation, in some outcrops, of both limbs of a first-generation fold being bent over the crest of a later fold; at the same time the earlier axial-plane cleavage is folded into an anticline or syncline. These particular folds may be a localized expression of the same movement responsible for the steeply plunging second-generation folds of the eastern part of the area. They may, conceivably, result from the movements which accompanied the emplacement of the Lebanon pluton.

Chevron folds of two types have been noted. In the more competent gently folded beds near the center of the Plainfield syncline, crinkling of the rocks is accomplished by slippage along joint surfaces (Fig. 10a); this chevron folding is evidently due to the failure of the rock, because of its rigidity, to yield by simple flexure or flow. German geologists have applied the term "*Totfaltung*" (dead folds) to these structures, the inference being that chevron folding occurs only in those rocks which have lost their intergranular plasticity and cannot yield by flowage or flexure. If this concept is correct, the chevron folds of figure 10a may represent a refolding of the rocks subsequent to their metamorphism.

The other type of chevron folding is a crinkling of steeply dipping schistosity along horizontal or gently westward-dipping slip planes. Along many of these slip planes the upper block has slid westerly, but this is not consistently so. These chevron folds are the youngest structures in the rocks, have a wave length seldom exceeding a few inches, and are found chiefly in the rocks immediately west of the Ammonoosuc and Monroe thrust faults. One possible origin is drag along a fault, which would tend to produce a fold with a subhorizontal or westerly dipping axial plane (Fig. 10b). There are not sufficiently good vertical exposures along either fault to prove the relations diagrammatically illustrated by the sketch, but it is significant that the chevron folds are best developed in those localities where the cleavage undergoes fairly abrupt changes in dip, or rolls from an easterly to a westerly dip. Another suggestion (Billings, verbal communication) is that the slip planes are a series of fractures developed as shear planes complementary to a main shear direction (*i.e.*, the major fault plane). Such complementary shears would have the observed westerly displacement of the upper block with respect to the lower one.

Faults

NORTHEY HILL THRUST: The Northey Hill thrust fault, which Chapman (1939, Pl. 1) has traced to the southeastern edge of the Hanover quadrangle, brings Orfordville schists upward on the west into contact with Littleton schists on the east. The projected course of the fault through the Hanover quadrangle is in an area where exposures are poor, and the fault cannot

be detected. Because metamorphic isograds are not offset by the fault in the Littleton-Moosilauke area Billings (1937, p. 531) has concluded that it is pre-metamorphic.

AMMONOOSUC THRUST: The only well-marked fault within the Hanover quadrangle is the post-metamorphic Ammonoosuc thrust. Fault breccia, mylonite, silicified or pyritized zones, slickensided surfaces, drag folds, and an abrupt juxtaposition of low-on middle-grade metamorphic rocks provide evidence of the fault from the northern to the southern boundary of the area. The general strike of the fault is northerly, along the Connecticut River, and the dip is between 35°–50° W. Displacement as deduced from the slickensides has been directly up the dip, and the hanging wall has moved eastward several thousand feet. There is no way of computing net slip in this area, but Billings (1937, p. 529) has estimated it between 14,000 and 18,000 feet in the Littleton-Moosilauke area, where the geologic relations indicate that this is a shear thrust and that younger rocks have been thrust over older ones. The total mapped length of the Ammonoosuc fault from the northeastern part of the Littleton quadrangle to the southern part of the Hanover quadrangle is more than 70 miles.

Field relations at the south end of the Ammonoosuc thrust are complex (Pl. 1, section C-C′ and D-D′), and a large plate of the Hardy Hill quartzite and related rocks is included as a slice in the thrust zone. The Post Pond volcanics are eliminated between the quartzite and the Gile Mountain formation on the west (section D-D′) because the Ammonoosuc fault has cut across an anticline (mapped by Chapman in the Claremont quadrangle) and has moved younger rocks eastward over older ones. The relations here are analogous to those in the Littleton-Moosilauke area discussed by Billings (1937, p. 529) at some length.

MONROE FAULT AND CORRELATION OF "NEW HAMPSHIRE" AND "VERMONT" STRATIGRAPHIC SEQUENCES: The Monroe fault (Eric, White, and Hadley, 1941, p. 1900) is an important structural boundary because it is the contact between the "New Hampshire" and "Vermont" stratigraphic sequences. In areas to the north the fault is indicated by the elimination of stratigraphic units along the eastern side of the fault, the presence of breccia in the Littleton quadrangle, an offset in metamorphic isograds in the Woodsville quadrangle (White and Billings, 1951, Pl. 1), and the truncation of beds in the Littleton quadrangle (Eric, 1942, Pl. 1) White and Billings (1951, p. 670) indicate it has a mapped length of at least 72 miles. They estimate the throw in the Woodsville quadrangle to be at least 5000 feet, with the west side overriding the east. Dip is steep.

The only structural indication of the Monroe fault in the Hanover quadrangle is the somewhat questionable evidence of chevron folding in some of the rocks west of the inferred position of the fault, a very slight divergence in the strike of the formations on either side of the projected trace of the fault, and the fact that, in the Strafford quadrangle, stratigraphic units on either side of the fault are lithologically different from analogous units in the Hanover quadrangle. The following suggest that the fault probably dies out somewhere in the Hanover quadrangle: (1) it has not been found in the quadrangles to the south, (2) there are at least two localities a mile south of the village of Norwich, Vt., where one can traverse practically continuous outcrops from the Post Pond volcanics, into a series of "pin stripe" schists which form the easternmost part of the Meetinghouse slate, and (3) the writer found no exposure of the elusive fault, despite the fact that he walked out the critical contact for 12 miles. Its southern terminus has been indicated, on inadequate evidence, somewhere in the vicinity of the Ottauquechee River (Pl. 1). Thus the Gile Mountain and Orfordville formations are presumably, over at least part of their boundary, in sedimentary contact.

A great number of cleavage-bedding and drag-fold readings (Pl. 2) near the critical contact indicate that the Gile Mountain formation overlies the Orfordville. Moreover, 1 mile north of the village of Norwich, Vt., are exposures of pillow lava in the Post Pond member of the Orfordville formation almost at its contact with the Gile Mountain beds; these pillows likewise indicate that the Gile Mountain beds overlie the Orfordville, unless the unrecognized Monroe fault lies between the outcrops of the two formations. If Gile Mountain rocks overlie Orfordville, the Orfordville and Waits River

formations are probably equivalent, and different facies of the same formation; the Post Pond and Standing Pond volcanics are the same bed; the Albee and Gile Mountain formations are identical; the Meetinghouse slate is a basal member of the Gile Mountain formation and grades into quartzites toward the west. Geologic relations in the Claremont quadrangle, although not yet fully unravelled, fail to demonstrate the equivalence of the Post Pond and Standing Pond members.

There are several reasons, moreover, for doubting the above correlations and for not placing complete confidence in the drag folds (which may partly be second-deformation structures) or in the pillow lavas (which are possibly separated from the rocks to the west by the Monroe fault). If the above correlation were correct, the Meetinghouse slate would be the basal part of the Gile Mountain formation. The basal portion of the Albee formation, where exposed in New Hampshire, is a quartzite. It is evident that the Meeting house slate could not be an equivalent of the basal Albee formation because the two units are distinctively different lithologically where they adjoin along the Monroe fault in the Mt. Cube quadrangle (Hadley, 1950, Pl. 1). Eric and Billings (verbal communication), who have studied the Albee and Gile Mountain formations extensively, are likewise convinced that the two formations are not equivalent. Likewise, the writer knows of no good evidence that the Orfordville formation grades westward into the Waits River, and of no locality where this facies change may be demonstrated. There is, as yet, no proof of the equivalence of the Standing Pond and Post Pond amphibolites.

In the face of this negative evidence one is left with the alternative that the Orfordville formation is younger than the Gile Mountain or partially equivalent to it. Some evidence favors this correlation. In the northern part of the Hanover quadrangle the lower Orfordville formation and the Meetinghouse slate are black phyllites or schists. The writer cannot distinguish these two units, where they adjoin along the Monroe fault in the Mt. Cube quadrangle. Both units become increasingly arenaceous in the southern part of the Hanover quadrangle and in that locality appear to be lithologically identical. If the Monroe fault does not persist throughout the Hanover quadrangle, the Meetinghouse slate and the Post Pond volcanics are in direct sedimentary contact. Geologic relations, though complex, suggest that at the northern end of the Plainfield syncline (Pl. 2) the Meetinghouse slate plunges beneath the Post Pond volcanics. This would mean that the Meetinghouse slate is the equivalent of the part of the Orfordville formation below the Post Pond volcanics and that the "New Hampshire sequence" overlies the "Vermont sequence". The writer favors this correlation. It is essentially the same as that adopted by Kruger (1946, p. 169) and Moore (1949, p. 1617) in southern New Hampshire and Vermont, although in those areas the Gile Mountain formation was not recognized, probably because it wedges out and grades into the Waits River formation. Meta-quartz conglomerates, however, have not been mapped within the Meetinghouse slate, although the Hardy Hill quartzite is a distinctive member of the lower Orfordville formation. The absence of a meta-quartz conglomerate in the Meetinghouse slate is possibly due to the fact that the Hardy Hill quartzite is discontinuous and may lense out toward the west.

Cleavage

Axial-plane, bedding, slip, and fracture cleavage are all represented in the folded rocks of the Hanover area. Most outcrops show essential parallelism of cleavage and bedding, but in many cases examination of near-by exposures indicates that the dominant cleavage is of the axial-plane type, and that isoclinal folding accounts for the parallelism of cleavage and bedding. Throughout most of the southwestern ninth of the quadrangle, for example, bedding and cleavage are parallel and are warped into a series of gentle anticlines and synclines (Pl. 2). Only on a few cliff faces can it be seen that the Waits River formation has been thrown into a series of essentially horizontal isoclinal recumbent folds, and that the anticlines and synclines shown on Plate 2 represent a warping of these earlier structures. In the broad area between the Pomfred and Hanover domes, bedding and cleavage in the

Gile Mountain formation are, likewise, essentially parallel, and in relatively few outcrops does cleavage cut the bedding at the noses of folds.

Axial-plane cleavage is evidently developed in most argillaceous or calcareous metamorphic rocks, and the parallelism of bedding and cleavage in these rocks is undoubtedly due to isoclinal folding. In more competent rocks, however, such as quartzites, quartz-mica schists, and amphibolites, bedding cleavage appears to be the dominant structure, and in many places bedding- and axial-plane cleavage cannot be differentiated. In the folds pictured in Figures 6, 9, and 10a, all the competent beds are foliated parallel to the bedding plane; this foliation wraps around the noses of the folds despite the development of axial-plane cleavage in the phyllitic rocks of Figure 9 and the calcareous rocks of Figure 6. Such bedding cleavage is doubtless partly mimetic in origin, but its development is probably aided to some extent by flexural slippage parallel to bedding planes. Because of uncertainty as to the extent of the development of bedding- versus axial-plane cleavage, particularly within the competent Gile Mountain formation, the recognition of second-deformation features in these rocks presents a formidable problem. White (1949, p. 590) and Hadley (1950, p. 28–32) have identified second-deformation folds in the area to the north partly on the basis of cleavage wrapping around the noses of the younger folds, and the frequent presence of axial-plane slip cleavage in these structures. Similar structures occur, however, over large areas of the Meriden anticline and the Plainfield syncline, both of which belong to a series of northeasterly-trending folds classified by White (1949, p. 590) as early-deformation structures. Unless, therefore, slip cleavage transects recognizable axial-plane cleavage, the recognition of a second-deformation structure on the basis of a fold in cleavage and of slip cleavage in the axial plane of such a fold is an extremely fallible field criterion.

Slip cleavage differs from fracture cleavage in its orientation and in this area parallels the axial planes of folds. As seen either in thin sections or in outcrop, slip cleavage develops along closely spaced micro-faults; micas belonging to bedding- or axial-plane cleavages are characteristically dragged into parallelism with the slippage planes. Slip cleavage such as this is mechanically more similar to axial plane than to fracture cleavage (White, 1949, p. 591–593), as there is appreciable elongation of the rock parallel to the cleavage plane and thinning normal thereto. Although some slip cleavage is clearly developed along second-deformation structures, such as the minor folds north of Plainfield, N. H., and the reverse drag folds west of Hartford, Vt., its presence is not a reliable criterion of a second-deformation structure. In the quadrangles to the north, White (1949, p. 591–593) and Hadley (1950, p. 28–32) have been able to map a progressive westward increase in the intensity of slip cleavage in the direction of the cleavage domes, but a similar situation is not apparent in the rocks of the Hanover area.

Fracture cleavage, here understood to be a rock cleavage developed at some angle to the axial plane of folds and ordinarily characterized by a lack of mineral parallelism, is a fairly rare type of cleavage. It is prominently developed only at two localities in the area—at Fernald and Smith Hills in the southeastern ninth of the quadrangle (Pl. 1), and near the mouth of the Ottauquechee River. In both places it occurs only in the competent beds, and there is microfaulting along the cleavage planes. The considerable amount of rock flowage which most rocks have undergone doubtless accounts for the rarity of this type of cleavage.

Lineation

Within the metamorphic rocks lineation is generally conspicuous and appears in a variety of ways. Axes of major and minor folds, crinkles, cleavage-bedding or cleavage-cleavage intersections, parallel orientation of elongate minerals and mineral streaks, deformed pebbles and amygdules, boudinage, and slickensiding are all present in varying degrees of development. Fold axes and cleavage-bedding intersections or crinklings are the most conspicuous lineations and boudinage, mineral lineation, and elongate pebbles or amygdules generally parallel these structures. The linear structures (Pl. 2) are more strongly developed around the three major anticlines of the quadrangle.

Around the Lebanon "dome" there is a strong mineral lineation in the Post Pond volcanics, and a crinkling in the schists parallel the linear structures within the pluton. This lineation has been impressed on all rocks for approximately 5 miles from the contact of the "granitic" rocks and reflects the movement of a great block of the earth upward and southeastward during the emplacement of the Lebanon pluton. Calculations of the degree of extension of the rocks parallel to the lineation, based on measurements of deformed amygdules (*cf.* Cloos, 1947, p. 861–865), indicates elongation of as much as 96 per cent. A somewhat anomalous situation arises at the south end of the dome, where the minor fold axes parallel the lineation and, therefore, the extension direction. According to generally accepted petrofabric terminology, a fold axis is *b*, (Fairbairn, 1949, p. 6), but here it is more properly considered as *a* because it parallels the major movement in the rocks. This point is made because it has a bearing on the petrofabrics of these rocks.

Lineations south of the Lebanon dome pass through a transition into those which reflect the structure of the Meriden anticline. Some fold axes and crinklings on the eastern side of the latter structure have a steep plunge and are due to the late-deformation east-side-north movement. Most of the cleavage-bedding lineations and the minor folds, however, plunge gently north or south in accordance with the plunge of the major fold. Elongated quartz pebbles in the Hardy Hill quartzite (with an average major axis of approximately 4 cm) are elongated downdip and flattened in the bedding plane on the flanks of the anticline; on the well-exposed northern nose the long dimensions of the pebbles parallel the fold axis. Their average extension parallel to the axis is 75 per cent, but in the weakly foliated bedding plane they have undergone very little distortion (−6%) and have been flattened normal to the bedding by 37 per cent. Arcuation of the axis of the Meriden anticline is evidently not the explanation of the distortion recorded by these elipsoidal pebbles. However, slippage along the bedding planes in a direction normal to the fold axes, combined with arcuation, may have produced the triaxial strain necessary for this puzzling distortion. In other recorded examples of pebbles in conglomerates elongated parallel to fold axes (Fairbairn, 1936, p. 670–680; Runner, 1934, p. 362), the intermediate pebble axis lies in the axial plane of the fold rather than in the bedding, and the extension has been attributed to shearing normal to the fold axis.

Within the Pomfret dome most minor fold axes, regardless of the intensity of folding, plunge either north or south, although a few folds on the eastern and western flanks plunge diagonally down the dip. Hornblende lineation in the Standing Pond amphibolite parallels fold axes. On the flanks of the major structure it is essentially horizontal. Crinkling in the phyllitic rocks parallels the dip direction, however, and therefore the *a* fabric axis.

Petrofabrics

An extensive petrofabric study of the rocks of this area has not been undertaken, although certain problems, such as the distortion of conglomerate pebbles, and the possibility of the recognition of superimposed deformations merit further investigation. Kaiser (1938, p. 129–133) has published a series of petrofabric diagrams of quartz and biotite in three rocks within and peripheral to the Lebanon pluton. All of these show girdles essentially normal to the prominent lineation (Pl. 3). Quartz girdles similar to Kaiser's have been determined by the writer in three specimens from the area west of the pluton, one of them 3 miles from the granite contact. Unlike Kaiser's diagrams, these showed no minor point concentration in the plane of foliation, and normal to the lineation, nor did it appear that the girdles were slightly offset from normality to the lineation. These differences, however, are minor, and the presence of quartz girdles normal or nearly normal to the lineation for rocks both inside and outside the pluton is the outstanding feature of the rock fabric.

As with a great many other petrofabric diagrams, the significance of the girdle is debatable. Kaiser's (1938, p. 129–134) interpretation, following conventional lines, is that the girdle pattern is that of a typical tectonite, and that the rocks have undergone intergranular rotation in a plane normal to the lineation; the point maximum in the girdle is due to shear.

The linear direction is therefore *b*, and the direction normal to it (coinciding with his point maximum) is *a*. On the other hand, if the writer's interpretation of the Lebanon dome is correct, the lineation parallels the direction of major rock movement, and therefore lies in the *a* fabric direction. Certain of the minor folds, particularly those at the south end of the Lebanon dome, would likewise parallel this axis, and the quartz girdles would be *bc* girdles. Under this set-up the fabric diagrams would be in accord with Anderson's (1948, p. 99–132) hypothesis that the direction of shearing or transport must be normal to the plane of the girdles. The field facts, insofar as the writer has been able to interpret them, favor this thesis. Balk's (1952, p. 415–435) recent demonstration of *bc* girdles in quartzites near thrust planes likewise accords with the idea that this is the normal fabric under strong unidirectional stress.

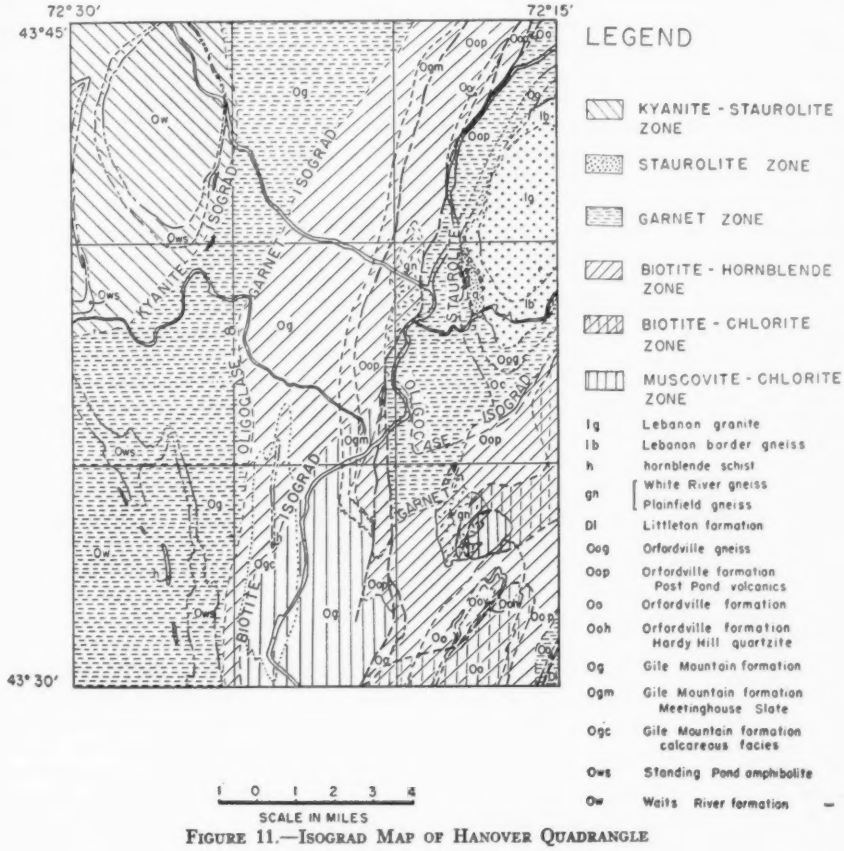


FIGURE 11.—ISOGRAD MAP OF HANOVER QUADRANGLE

METAMORPHISM

Introduction

The following comments on the metamorphism of the rocks of the Hanover area are based on field observation and the study of more than 400 hand specimens and 200 thin sections. Approximately 230 minerals have been studied optically to determine significant molecular proportions. Accuracy of the determinations of refractive index is ± 0.003 except in the case of the feldspars, where a set of oils graduated in steps of 0.002 was used.

The mineralogical composition of most rocks indicates that metamorphic equilibrium was attained. Small amounts of chlorite or sericite in some of the higher-grade rocks represent disequilibrium, but these minerals are not volumetrically nor areally important, and in most cases are probably retrogressive products formed under decreasing temperature conditions. A few rocks have been hydrothermally altered subsequent to the regional metamorphism, but these are not widespread, nor is their alteration extensive. Turner's (1948, p. 45–107) diagrams and terminology have been used in classifying the rocks into their appropriate facies, but isograds are shown on the metamorphic map (Fig. 11) because they are more easily recognized and used in the field.

Isograds and Cause of Metamorphism

The isograd map of the quadrangle (Fig. 11) indicates clearly the direct relation of the P-T conditions at the time of metamorphism to the two major domal structures of the area. The concentric distribution of the isograds about these structures and the geologic relations previously described leave little doubt that the essential cause of the metamorphic zoning was the upward projection of two masses of hot, low-density rock into a series of deeply buried and highly deformed sediments and volcanics.

The succession of metamorphic isograds is approximately that which has been noted in numerous similar areas, but a few peculiarities are worth noting. Staurolite is so rare in rocks of the western part of the area that an isograd cannot be drawn for it. The few rocks which contain staurolite also contain kyanite, and the isograds of the two minerals probably overlap. All the staurolite-bearing rocks are also characterized by the association biotite-muscovite-garnet-staurolite-kyanite but are probably not in disequilibrium (Barth, 1936, p. 822).

Garnet (almandine?) and oligoclase isograds are essentially coincident, but garnet is developed under lower P-T conditions in the metamorphosed volcanics than in the argillaceous rocks. This is contrary to the recently described pattern in the Black Hills, where Noble and Harder (1948, p. 967) have found that in the intrusive amphibolites higher P-T conditions are necessary for the appearance of garnet than are necessary in the pelitic schists. A possible explanation for the apparently anomalous situation in the Hanover area is that the garnets are manganiferous; this is to be expected in rocks that were originally, in part, submarine extrusives.

In the rocks peripheral to the border gneiss of the Lebanon granite, which were originally aluminous schists of the Orfordville formation, sillimanite is absent, and staurolite and kyanite are rare. The absence of sillimanite is due, in part at least, to the relatively low temperature of the Lebanon pluton, but the absence of kyanite and staurolite is more probably a compositional effect caused by metasomatic addition of lime, soda, and potash to the aluminous rocks. The net result of the metasomatism was to move the bulk composition of the wall rocks from fields where kyanite and staurolite were stable to those where micas, epidote, plagioclase, garnet, and microcline were the stable minerals. Both the granite and the bordering schists crystallized under P-T conditions of the staurolite-kyanite subfacies of the amphibolite facies (Turner, 1948, p. 81–85). Most of the mineralogical criteria, recently described by Ch'ih (1950, p. 926–933) as evidences of the granitization of the Wissahickon schist near Philadelphia, are repeated in the Lebanon granite and its surrounding schists. Worth noting, as examples, are: (1) the gradual decrease in muscovite and the increase in microcline toward the granite, (2) the presence of green biotite in the granite and brown biotite in the schists, (3) the abundance of symplectic intergrowths, and (4) the relative abundance of epidote and sphene with biotite near the edges of the pluton. It is evident that the granitization in the Philadelphia area, which is quite possibly identical in age with that in New Hampshire, has proceeded along a practically identical path.

Most of the critical isograd minerals are porphyroblastic, and textural relations show that their growth post-dated the deformation of the rocks. The peak of thermal metamorphism, therefore, occurred after the period of major rock deformation. S-shaped trains of inclusions in a few garnet porphyroblasts demonstrate some overlapping of the thermal and

deformational episodes, but such rotated porphyroblasts are relatively uncommon. More significantly, no porphyroblasts are ruptured or strained, whereas groundmass minerals particularly the micas, are in many places highly contorted. In some rocks plagioclase porphyroblasts lie athwart such layers of mica, and the relict contorted structure is easily traceable through the undistorted and homogeneous feldspar. Therefore, the metamorphism of these rocks probably proceeded in two essentially distinct stages. In the early stage of dynamic metamorphism temperature was relatively low and directed pressure fairly high; it is doubtful that rocks of higher metamorphic intensity than the greenschist facies developed during this phase. The later phase, which was essentially thermal, was directly related to the rise of the Lebanon and Pomfret domes. Most porphyroblastic minerals show no preferred orientation, and the effects of differential stress were obviously minor. Some porphyroblastic minerals, such as hornblende, do show preferred orientation, but the crystals lie in bedding planes, rather than in the axial planes of folds, and have evidently grown by mimetic crystallization.

Retrogressive Metamorphism

Retrogression of hornblende and garnet schists to lower-grade rocks in the foot wall of the Ammonoosuc thrust has been described by Hadley (1942, p. 171–173) and ascribed by him to a combination of shearing stress and low-temperature hydrothermal activity. This retrogression is likewise common in many of the foot-wall rocks of the Ammonoosuc thrust in the Hanover quadrangle, but the width of the retrogressed zone is in places, only a few feet; it nowhere approaches the 1000-foot average width which it maintains in the Mt. Cube quadrangle (Hadley, 1942, p. 171–173).

Widespread but mild retrogression of the schists throughout the eastern half of the Hanover quadrangle is marked by chloritization epidotization, and sericitization of all the rocks. Because this type of alteration also affects the Mississippian (?) diabase dikes it must be distinctly later than the regional metamorphism and is probably a hydrothermal process. The introduction of sulfide veins and replacements in the Lebanon granite and surrounding rocks appears to be genetically related to the hydrothermal metamorphism. Certain puzzling disequilibrium rocks of the western slide of the Mascoma quadrangle (consisting of calcite-chlorite-muscovite-andesine assemblages) possibly owe their origin to the same process.

Many of the rocks of the western part of the area have retrogressive chlorite and sericite in small quantities. There is no indication, however, that the rocks have undergone hydrothermal alteration, and their slight retrogression is attributed to falling temperatures in the closing stages of the regional metamorphism.

Special Features

In view of the well-known high concentration of boron in many shales, the presence of minor amounts of tourmaline in pelitic schists has long been regarded as an unreliable indicator of boron metasomatism. That such a type of metasomatism has affected rocks of the Hanover area, however, is indicated in some of the calcareous schists and amphibolites by sporadic concentrations of as much as 5 per cent of tourmaline. These concentrations are understandably rare, and they show no areal pattern with respect ro the regional isograds. According to Currier's (1947, p. 77) concept of regional granitization in New England such tourmalinization is normal manifestation, in the lower-grade rocks, of a broad "invasive metamorphism" whose end result, in eastern New England, has been the formation of granitized rocks.

Quartz veins penetrate all the metamorphics, and, because they are as abundant in the amphibolites and calcareous schists as in the arenaceous schists, there is little reason for supposing that they may have been generated *in situ* by metamorphic differentiation. Some veins were at least partially formed by replacement; opposite walls of the veins fail to correspond in their shapes, indicating a nondilatational behavior of the rock at the time the quartz was introduced. The quartz veins belong to the general period of metamorphism because they are occasionally sheared off and brecciated, and because they do not penetrate the Mississippian (?) diabases, even along joint surfaces.

Mineralogy of Metamorphic Rocks

PLAGIOCLASE-EPIDOTE: Figure 12 has been constructed largely as a check on Ramberg's theoretical plagioclase-epidote equilibrium diagram (1944, p. 51–53) in an effort to determine how closely its predicted relations are followed in the rocks of the Hanover area. The determination of the An content of the plagioclases was made with a set of index oils calibrated to ±0.002, and the feldspar compositions represented on the graph are believed to be accurate to within ±2 per cent of the An content. Estimates of the Ca/Ca + Na ratios of the epidote-plagioclase associations, on the other hand, are based on estimations of the relative proportions

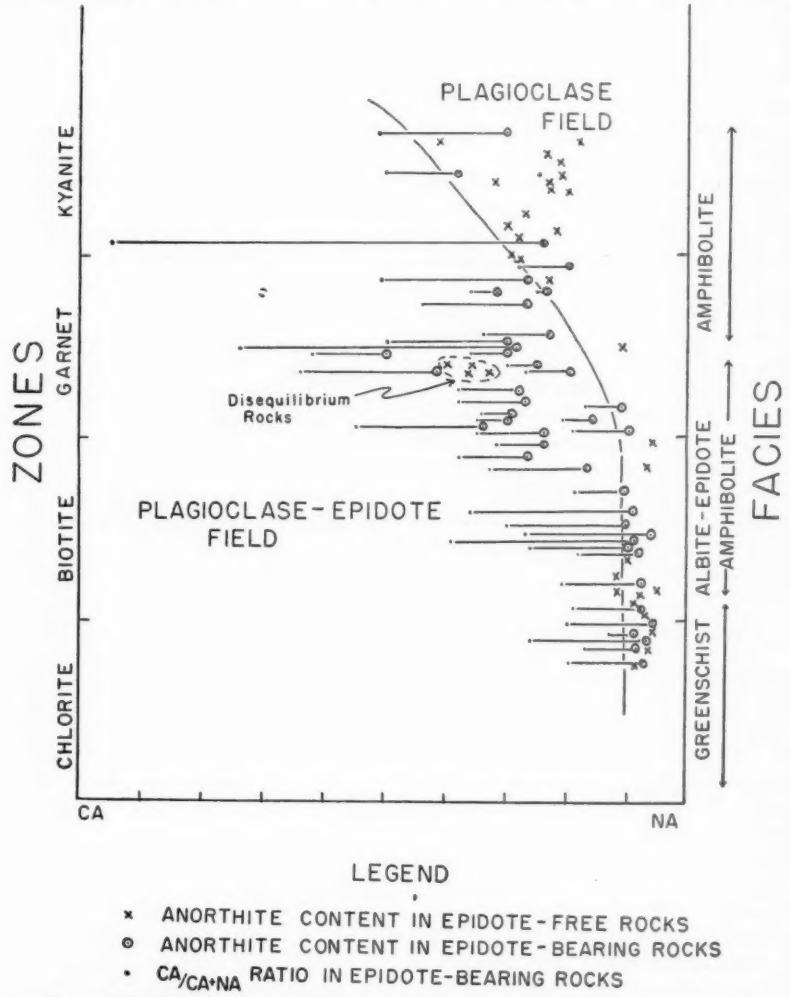


FIGURE 12.—PLAGIOCLASE AND PLAGIOCLASE-EPIDOTE ASSOCIATIONS IN ROCKS OF HANOVER AREA

Circles = Anorthite content of plagioclase in epidote-bearing rocks
Dots = Atomic ratio, Ca/Ca + Na, of rocks containing both epidote and plagioclase
Crosses = Anorthite content of plagioclase in plagioclase-free rocks

of epidote and plagioclase in the rock, plus the determined plagioclase composition. These points are, therefore, less accurately determined, but are less significant than the plagioclase compositions. Seventy-five rocks, of variable composition but of known location with respect to the regional isograds, were used in this study; all the rocks appeared to be in metamorphic equilibrium.

In a general way the separation of the plagioclase-epidote and plagioclase fields follows Ramberg's curve, and the data clearly indicate a gradual increase in the anorthite content of the plagioclases with increasing metamorphism, provided that calcium is available in the rock. On the other hand there are deviations of as much as 30 per cent An from the predicted values, and it is clear that anorthite content of a plagioclase in equilibrium with epidote is not, by itself, a completely satisfactory index of the intensity of metamorphism. Deviations from the equilibrium curve (Ramberg, 1944, p. 53) may be due to such factors as the presence of calcite, variations in the relative abundance of potassium, aluminum and water, or fluctuations in the vapor pressure of H_2O. Of these predicted deviating factors, the presence of calcite appears to be most important if that mineral is an important rock-forming constituent. Highly calcareous rocks of the Hanover region, especially within the albite-epidote-amphibolite facies, have much more calcic plagioclases than should be expected, although according to Ramberg (1944, p. 54) at high metamorphic P-T conditions the presence of calcite favors the formation of a sodic plagioclase. As far as could be determined, the relative abundance of micas in these rocks has little effect on the anorthite content of the plagioclase. The highest anorthite content determined (An_{50}) was in a quartz-mica schist of the amphibolite facies, lying within the garnet zone, but below the kyanite zone.

Other features brought out by the diagram are well known. There is, for example, a great paucity of metamorphic plagioclase in the range of An_{10}–An_{20}. The first appearance of oligoclase and more calcic feldspars coincides approximately with that of garnet (presumably almandine).

Epidote in equilibrium with the plagioclases is clinozoisite, some of it with up to 20 mol per cent of the iron epidote end member in solid solution. There is, however, no relation between the iron content of the epidote and the degree of metamorphism. As would be expected from the shape of the curve in Figure 11, clinozoisite in rocks of the kyanite and higher isograd zones is extremely rare.

AMPHIBOLES: Factors controlling the origin and chemical composition of calciferous amphiboles in metamorphic rocks are, at best, imperfectly understood. On both empirical and theoretical (Ramberg, 1952, p. 344–349) grounds, however, it is to be expected that with increasing P-T conditions there will be a progressive increase in both the amount of Al substituting for Si in the amphibole, and in the Fe/Fe + Mg atomic ratio. Ramberg (1949, p. 39–47) has demonstrated an equilibrium relation between hornblende and co-existing orthorhombic pyroxene in rocks of the granulite facies, and an apparent increase in the Fe contents of both minerals with increasing metamorphism. Wiseman (1934, p. 354–417) and Tilley (1938, p. 497–511) have shown in the epidiorites of the Scottish Highlands that the indices of refraction of chlorites and actinolitic amphiboles from the same specimens vary sympathetically with one another, indicating that the two minerals have proportionate Fe/Fe + Mg atomic ratios. To the writer's knowledge, these three papers report the only demonstrated systematic variations in the calciferous amphiboles of metamorphic origin. It is noteworthy that very little is known of the causes for the variability of common hornblende in the amphibolite facies, except for the general factor of rock composition.

Foslie (1945, p. 74–98) has published a curve for the hornblendes of the epidote-amphibolite facies (Ramberg, 1944, p. 71–72; Barth, 1952, p. 337–341) in which the mean index of refraction $\frac{(\alpha + \beta + \gamma)}{3}$ is plotted against the Niggli mg-number

$$\left(\frac{\text{MgO}}{\text{mol } 2Fe_2O_3 + \text{FeO} + \text{MnO} + \text{MgO}}\right).$$

This latter number is, of course, the same as the atomic ratio $\frac{\text{Mg}}{\text{Mg} + \text{Fe} + \text{Mn}}$. Foslie's curve is

based upon amphibole analyses recalculated from rock analyses, but seems to be valid for such analyzed hornblendes as that from the Littleton quadrangel (Billings, 1937, p. 556). The range of the P-T conditions of the *epidote-amphibolite* facies covers most of the occurrences of hornblende in the rocks of the Hanover area.

Application of the Foslie curve to 34 hornblendes from the Hanover quadrangle indicates that the *mg-number* ranges from 35 to 65, but there is no apparent correlation of this number with the degree of metamorphic intensity. None of the other observable mineralogic features of the rocks, such as the presence or absence of such minerals as epidote, garnet, and chlorite (the three minerals most commonly used in equations written to explain the genesis of hornblende), has any evident effect on the hornblende composition. Moreover, in contradistinction to the results of Tilley (1938) and Wiseman (1934), there is no correlation of the indices of refraction of hornblendes and chlorites coexisting in the same rock (10 samples). Reasons, therefore, for the compositional variations in the amphiboles of the Hanover area remain enigmatic.

Actinolite, some of it in homoaxial intergrowths with green hornblende, is recognizable in some of the rocks of the staurolite-kyanite subfacies of the amphibolite facies in this area. Similar actinolite has been mapped as an isograd mineral in the Woodsville quadrangle (White and Billings, 1951, Pl. 1), and lies on the high P-T side of the kyanite isograd. In both areas the actinolite is developed in the Waits River formation. In the Hanover area it is never developed where the rocks are highly feldspathic or rich in clinozoisite or mica. Its growth, therefore, under elevated P-T conditions is probably a reflection of a low aluminum content in certain rocks. Sundius (1946, p. 33) has discussed the possibility of simultaneous formation of tremolite-actinolite and of hornblende and has also noted the possibility of limited immiscibility between the two.

MICAS: Biotite, some of it porphyroblastic, is the commonest mica in the metamorphic rocks. Its β index of refraction in 30 rock specimens ranged from 1.608 to 1.658 but showed no evident relation either to the petrographic nature of the host rock or to the intensity of metamorphism. Kruger, who has determined the β index on 150 biotites of middle- and high-grade rocks (1946, p. 199), notes that "the indices of refraction from the middle- and high-grade zones are not significantly different." His finding that the index of refraction of the biotite in the lime-silicate rocks is low is, as expected, confirmed in this area. Phlogopitic mica is found only in the Waits River formation, and even there it is relatively rare. Phlogopite is probably produced in these rocks by the reaction of dolomite and muscovite; the rarity of phlogopite in the Waits River formation indicates that it was originally nondolomitic.

CARBONATES: Many of the calcareous rocks, particularly the Waits River formation, contain calcite or ankerite or both minerals. In 34 carbonates whose indices of refraction were determined the mol per cent of $MgCO_3$, as determined from Kennedy's (1947, p. 569) curves, ranged from 4 per cent to 10 per cent, but there was no relation between the Mg content and the metamorphic intensity. The Mg/Mg + Fe ratio of the ankerites ranged up to approximately 22 per cent, but it likewise showed no relation to the degree of metamorphism.

The reason for the presence of ankerite is not apparent. Calcite and magnetite have been observed in rocks from the lowest to the highest grade of metamorphism in this area; it is evident from their association that these two minerals do not react to form the ankerite. Ankerite is common in those rocks that contain hornblende, and it is suggested that, when garnet or chlorite react with clinozoisite to form hornblende, some of the Mg and Fe freed in that reaction may combine with calcite to produce ankerite. In some rocks, however, ankerite is associated with retrogressive chlorite, and its Fe and Mg are apparently derived from the breakdown of biotite or garnet.

Tectonic History

Paleozoic sedimentation in the Hanover area probably commenced in earliest Cambrian time, but the older Cambro-Ordovician rocks are not exposed, and the record begins with the accumulation, in Waits River time, of a group of predominantly calcareous sediments. To-

ward the close of this interval, the eugeosynclinal environment was radically altered, and a group of arenaceous sediments (the Gile Mountain formation) was deposited above the Waits River beds. Volcanism marked this change in paleogeographic conditions, the volcanic rocks accumulating in a unit now known as the Standing Pond amphibolite.

Geosynclinal conditions during the accumulation of the Gile Mountain formation are not accurately known, but it seems that the Gile Mountain beds have a limited areal distribution and that they accumulated as a pocketlike deposit in east-central Vermont. At the boundary between the Hanover and Claremont quadrangles the formation thins abruptly, and the Gile Mountain beds probably interfinger with the argillaceous material in the lower part of the Orfordville formation. Certain peculiar quartz conglomerates of the Orfordville rocks are not found within the Gile Mountain formation, however, and the extent of the interfingering as well as the exact stratigraphic relations of the two formations are imperfectly known. The southward thinning of the Gile Mountain formation may possibly be due to the fact that positive areas, some of which were Precambrian structures, remained emergent, or almost emergent, during the geosynclinal sedimentation.

The Orfordville formation largely represents an accumulation of black muds and sands, with minor conglomeratic and volcanic material in the lower part of the formation. A thick series of volcanics which seems to have originated in the east-central part of the Hanover quadrangle is now preserved as the Post Pond volcanic member of the Orfordville formation. This member is the youngest exposed Ordovician rock within the quadrangle.

The Taconic orogeny affected these formations (Billings, 1937, p. 468), but its intensity in this area is impossible to decipher. Silurian and Devonian formations were deposited upon the folded Cambro-Ordovician rocks, and a small portion of the Devonian Littleton formation is now preserved along the east side of the Northey Hill thrust in the southeastern portion of the Hanover quadrangle; other Silurian and Devonian rocks have been removed from this area by erosion.

The Acadian orogeny at the close of the Devonian was probably (Billings, 1937, p. 463–566; White and Billings, 1951, p. 695) the major deformational episode in western New Hampshire and was accompanied by the intrusion of great quantities of granitic rock which were important agents of metamorphism. The synkinematically intruded Lebanon granite and the Pomfret cleavage dome are the local expressions of the upward rise of granitic material in the Acadian orogeny.

From the close of the Devonian to the present the geologic record is meager. Mississippian (?) diabase dikes, Pleistocene glacial deposits, and recent alluvium yield but scanty clues to geologic events from the Acadian orogeny to the present.

References Cited

Anderson, E. M., 1948, On lineation and petrofabric structure, and the shearing movement by which they have been produced: Geol. Soc. London Quart. Jour., 104, p. 99–132.

Balk, R, 1946, Gneiss dome at Shelburne Falls, Mass.: Geol. Soc. America Bull., v. 57, p. 125–160.

—— 1952, Fabric of quartzite near thrust planes: Jour. Geology, v. 60, p. 415–435.

Bannerman, H. M., 1941, Sillimanite, andalusite, kyanite, and mica schist deposits: N. H. Plan. and Develop. Com. Spec. Rept., pt. IV, 5 pages.

Barth, Tom. F. W., 1936, Structural and petrologic studies in Dutchess County, N. Y., Part II: Geol. Soc. America Bull., v. 47, p. 775–780.

——, 1952, Theoretical petrology: New York, John Wiley and Sons, Inc., 387 pages.

Bean, R. J., 1951, The relation of gravity anomalies to the geology of central Vermont and New Hampshire: Harvard Univ. Ph. D. Thesis.

Billings, M. P., 1937, Regional metamorphism of the Littleton-Moosilauke area, New Hampshire: Geol. Soc. America Bull., v. 48, p. 463–566.

Billings, M. P., and White, W. S., 1950, Metamorphosed mafic dikes of the Woodsville quadrangle, Vermont and New Hampshire: Am. Mineral., v. 35, p. 629–644.

Cady, W. M., 1950, Fossil cup corals from the metamorphic rocks of central Vermont: Am. Jour. Sci., v. 248, p. 488–497.

Chang, P. H., 1950, Structure and metamorphism of the Bridgewater-Woodstock area, Vermont: Harvard Univ. Ph. D. Thesis.

Chapman, C. A., 1939, Geology of the Mascoma quadrangle, New Hampshire: Geol. Soc. America Bull., v. 50, p. 127–180.

——, 1942, Intrusive domes of the Claremont-Newport area, New Hampshire: Geol. Soc. America Bull., v. 53, p. 889–916.

Ch'ih, C. S., 1950, Structural petrology of the Wissahickon schist near Philadelphia: Geol. Soc. America Bull., v. 61, p. 923–956.

Chidester, A. G., Billings, M. P., and Cady, W. M., 1951, Talc investigations in Vermont: U. S. Geol. Survey, Prelim. Rept. Circ. 65, 33 pages.

Clark, T. H., 1934, Structure and stratigraphy of southern Quebec: Geol. Soc. America Bull., v. 45, p. 1–20.

Cloos, E., 1947, Oolite deformation in the South Mountain fold, Maryland: Geol. Soc. America Bull., v. 58, p. 843–918.

Cooke, H. C., 1937, Thetford, Disraeli, and eastern half of the Warwick map-area, Quebec: Geol. Survey Canada Mem. 211, 160 pages.

Currier, L. W., 1947, Granitization and its significance as a regional metamorphic process in New England: Wash. Acad. Sci. Jour., v. 37, p. 75–86.

Currier, L. W., and Jahns, R. H., 1941, Ordovician stratigraphy of central Vermont: Geol. Soc. America Bull., v. 52, p. 1487–1512.

Doll, C. H., 1943 a, A Paleozoic revision in Vermont: Am. Jour. Sci., v. 241, p. 57–64.

——, 1943 b, A brachiopod from mica schist, South Strafford, Vermont: Am. Jour. Sci., v. 241, p. 676–679.

——, 1944, A preliminary report on the geology of the Strafford Quadrangle, Vermont: Vt. State Geologist 1943–1944, Rept. p. 14–28.

——, 1951, Geology of the Memphremagog quadrangle and the southeastern portion of the Irasburg quadrangle, Vermont: Vt. Geol. Survey Bull. 3, 80 pages.

Eric, J. H., 1942, Geology of the Vermont portion of the Littleton quadrangle: Harvard Univ. Ph.D. Thesis.

Eric, J. H., White, W. S., and Hadley, J. B., 1941, Monroe fault of New Hampshire and Vermont (Abstract): Geol. Soc. America Bull., v. 52, p. 1900.

Fairbairn, H. W., 1936, Elongation in deformed rocks: Jour. Geology, v. 44, p. 670–680.

——, 1949, Structural petrology of deformed rocks: Addison-Wesley Press, 344 pages.

Foslie, S., 1945, Hastingsites and amphiboles from the epidote-amphibolite facies: Norsk Geol. Tidsskrift, v. 25, p. 74–98.

Fowler-Billings, K., 1944, Igneous and metasedimentary dikes of the Mt. Washington area, New Hampshire: Geol. Soc. America Bull., v. 44, p. 1255–1278.

Gilluly, J., 1933, Replacement origin of the albite granite near Sparta, Oregon: U. S. Geol. Survey Prof. Paper 175, p. 65–81.

Goldthwait, J. W., 1910, Glacio-lacustrine and postglacial features of the Connecticut Valley near Hanover, N. H. (Abstract): Science, New Ser., v. 32, p. 126.

——, 1916, Evidence for and against the existence of local glaciers in Vermont: Vt. State Geologist Rept. 10, p. 42–73.

Goldthwait, J. W., L., and R. P., 1951, The geology of New Hampshire: Part I—Surficial Geology, N. H. State Plan. and Develop. Comm., Concord, N. H., 83 pages.

Hadley, J. B., 1942, Stratigraphy, structure, and petrology of the Mt. Cube area, New Hampshire: Geol. Soc. America Bull., v. 53, p. 113–176.

——, 1950, Geology of the Bradford-Thetford area, Orange County, Vermont: Vt. Geol. Survey Bull. 1, 36 pages.

Hallimond, A. F., 1943, On the graphical representation of the calciferous amphiboles: Am. Mineralogist, v. 28, p. 65–89.

Hitchcock, C. H., 1874; 1877; 1878, Geology of New Hampshire: 3 volumes and atlas. Concord, N. H., 2069 pages.

——, 1908, Geology of the Hanover, N. H. quadrangle: Vt. State Geologist Rept. 6, p. 139–186.

Hitchcock, E., 1861, Report on the geology of Vermont: 2 vols., Claremont Mfg. Co., 988 pages.

Jackson, C. T., 1844, Final report on the geology and mineralogy of the state of New Hampshire: Concord, N. H., 376 pages.

Johannsen, A., 1932, A descriptive petrography of the igneous rocks: 4 vols., Univ. Chicago Press, 1622 pages.

Kaiser, E. P., 1938, Geology of the Lebanon granite: Am. Jour. Sci., v. 37, p. 107–136.

Kennedy, G. C., 1947, Charts for correlation of optical properties with chemical composition of some common rock-forming minerals: Am. Mineralogist, v. 32, p. 561–573.

Kruger, F. C., 1946, Structure and metamorphism in the Bellows Falls quadrangle of New Hampshire and Vermont: Geol. Soc. America Bull., v. 57, p. 161–206.

Krumbein, W. C., and Sloss, L. L., 1951, Stratigraphy and sedimentation: W. H. Freeman and Co., San Francisco, 497 pages.

Lougee, R. J., 1935, Hanover submerged: Dartmouth Alumni Mag., v. 27, no. 8, p. 5–6.

Merritt, J. W., 1913, Structural and metamorphic geology of the Hanover district of New Hampshire: Vt. State Geologist 12th Ann. Rept., p. 1–36.

Moore, G. E., 1949, Structure and metamorphism of the Keene-Brattleboro area, New Hampshire-Vermont: Geol. Soc. America Bull., v. 60, p. 1613–1670.

Noble, J. F., and Harder, J. O., 1948, Stratigraphy and metamorphism in a part of the northern Black Hills and near the Homestake mine: Geol. Soc. America Bull., v. 59, p. 941–976.

Ramberg, H., 1944, Petrological significance of some subsolidus phase transitions in mixed crystals; Norsk Geol. Tidsskrift, v. 24, p. 42–73.

——, 1948, Radial diffusion and chemical stability in the gravitational field: Jour. Geology, v. 56, p. 448–458.

——, 1949, The facies classification: a clue to the origin of the quartzo-feldspathic rocks: Jour. Geology, v. 57, p. 18–54.

——, 1952, Chemical bonds and distribution of cations in silicates: Jour. Geology, v. 60, p. 331–355.

Richardson, C. H., 1906, Areal and economic geology of northwestern Vermont: Vt. State Geologist Rept. 5, p. 63–116.

Runner, J. J., 1934, Pre-Cambrian geology of the Nemo district, Black Hills, South Dakota: Am. Jour. Sci., 5th ser., v. 228, p. 353–372.

Sundius, N., 1946, The classification of the hornblendes: Sver. Geol. Undersoknung, Arsbok 40, no. 4, 35 pages.

Tilley, C. E., 1938, The status of hornblende in low-grade metamorphic zones of green schists: Geol. Mag., v. LXXV, p. 497–511.

Turner, F. J., 1948, Evolution of the metamorphic rocks: Geol. Soc. America Mem. 30, 342 pages.

White, W. S., 1949, Cleavage in east-central Vermont: Am. Geophys. Union Trans., v. 30, p. 587–594.

——, and Billings, M. P., 1951, Geology of the Woodsville quadrangle, Vermont-New Hampshire: Geol. Soc. America Bull., v. 62, p. 647–696.

White, W. S. and Jahns, R. H., 1950, Structure of central and east-central Vermont: Jour. Geology, v. 58, p. 179–220.

Wiseman, J. D. H., 1934, The Central and Southwest Highland epidiorites: a study in progressive metamorphism: Geol. Soc. London Quart. Jour., v. XC, p. 354–417.

Dartmouth College, Hanover, N. H.

Manuscript Received by the Secretary of the Society, January 5, 1953

Project Grant 507/47/48

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 66, PP. 147-148, 1 FIG. JANUARY 1955

Short Notes

THICKNESS OF BEDDING AND PARTING IN SEDIMENTARY ROCKS

By Henry H. Gray

Recent papers by McKee and Weir (1953) and Ingram (1953; 1954) have proposed definitions and quantification of terms commonly used in surface stratigraphy to describe bedding and parting in sedimentary rocks. Reaction to and acceptance of these terms will be varied, but the suggested classifications are significant advances toward standardization of terminology.

Table 1.—Thickness of Bedding and Parting Units in Sedimentary Rocks

Thickness	Bedding classes	Parting classes
More than 1 m	Very thick (bedded, cross-bedded, incline-bedded)	Very thick (blades, rods, plates, blocks)
10 cm–1 m	Thick (bedded, etc.)	Thick (blades, etc.)
1 cm–10 cm	Medium (bedded, etc.)	Medium (blades, etc.)
1 mm—1 cm	Thin (bedded, etc.)	Thin (blades, etc.)
Less than 1 mm	Very thin (bedded, etc.)	Very thin (blades, etc.)

Both systems are somewhat cumbersome, however. Simpler schemes for the description of bedding and parting thickness are shown in Table 1. These were developed and modified from a system proposed by Payne (1942). The five classes describe thickness only, not shape. The ambiguous and misused terms *massive*, *super*, and *extra* are avoided. Classes are logical and on a simple logarithmic scale, and the names are easily remembered.

Bed and layer are considered essentially synonymous with stratum as defined by McKee and Weir (1953); against the inadvisability of duplication of terms must be weighed the advisability of having synonyms available for variety of expression. Distinction of beds from laminae solely on the basis of thickness results in lateral equivalence of beds and laminae because all beds must some place thin to extinction. Terminology for the description of bedding properties other than thickness is well discussed by McKee and Weir (1953).

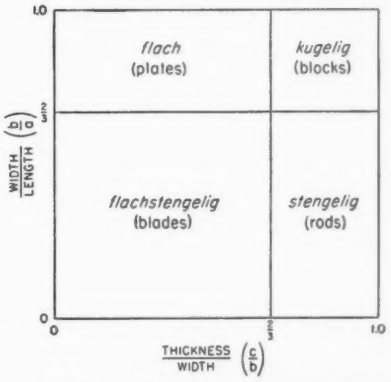


Figure 1.—Shape of Parting Fragments of Sedimentary Rocks
Modified from Zingg (1935)

Parting fragments are described in terms of shape and orientation as well as thickness. Zingg's (1953) classification of pebble shapes may readily be adapted to parting fragments (Fig. 1). Orientation of fragments is described as parallel or nonparallel to bedding.

References Cited

Ingram, R. L., 1953, Fissility of mudrocks: Geol. Soc. America Bull., v. 64, p. 869–878.

Ingram, R. L., 1954, Terminology for the thickness of stratification and parting units in sedimentary rocks: Geol. Soc. America Bull., v. 65, p. 937–938.

McKee, E. D., and Weir, G. W., 1953, Terminology for stratification and cross-stratification in sedimentary rocks: Geol. Soc. America Bull., v. 64, p. 381–390.

Payne, T. G., 1942, Stratigraphical analysis and environmental reconstruction: Am. Assoc. Petroleum Geologists Bull., v. 26, p. 1697–1770.

Zingg, Th., 1935, Beitrag zur Schotteranalyse: Schweiz. Min. u. Pet. Mitt., Bd. 15, p. 39–140.

INDIANA GEOLOGICAL SURVEY, BLOOMINGTON, INDIANA

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 66, PP. 149-152, 1 FIG. JANUARY 1955

PRELIMINARY REPORT ON THE SYSTEM SiO_2-H_2O

BY O. F. TUTTLE AND J. L. ENGLAND

Introduction

The water-rich portion of this system has been studied extensively in recent years (Morey and Hesselgesser, 1951; Kennedy, 1944; 1950; Van Nieuwenburg and Van Zon, 1935), and the solubility of silica in water is consequently well known; the maximum solubility reported is less than 1 per cent.

Despite intense interest, one of the most important parts of the system, the solubility of water in liquid silica, has not been studied. This report deals with a preliminary study of the high silica portion of the system. Unfortunately, the apparatus failed before an adequate check of reproducibility could be made. However, the results of the preliminary experiments are of general interest, and it is believed desirable to present them at this time despite the uncertainties, because the work cannot be continued in the immediate future.

Pressure measurements were made with bourdon gauges which were calibrated with a dead-weight gauge and are believed to be accurate to ±10 per cent. The temperature measurements are somewhat uncertain because the failure took place before a careful calibration of the thermocouple was made. However, it seems unlikely that the thermocouple would be in error by more than ±10°C.

Experimental Method

All experiments were carried out in platinum capsules made by sealing both ends of ¾-inch lengths of platinum tubing by welding with a carbon arc. Tubing having an internal diameter of 2.5 mm with a wall thickness of 0.2 mm is very satisfactory. Charges are made up as follows: One end of a ¾-inch piece of platinum tubing is pinched shut and welded along the straight edge produced by collapsing the tube. The tube is cleaned thoroughly and annealed at 1200°–1500°C for a few minutes. After annealing, the tube is weighed, and the desired amount of quartz, tridymite, or silica glass is placed in the tube which is then reweighed to determine the exact weight of the silica. Water is then added by a small capillary pipette; the exact amount is determined by reweighing. The tube is then sealed by welding as before; the water is prevented from escaping by pinching a tube in a small vise while welding. The jaws of the vise are narrow and do not pinch the portion of the tube holding the silica and water. After the final welding, the tube is reweighed as small amounts of platinum are sometimes lost during welding. At the end of the run the tube is weighed, and any gain or loss in weight can be readily detected.

All runs were made in an internally heated pressure vessel similar to that described by Goranson (1931). Pressure is generated by pumping carbon dioxide into the pressure vessel and is applied to the charge by collapse of the platinum tubing. It is not feasible to pump water directly into the pressure vessel because of the relatively high electrical conductivity of water under these conditions which would "short-out" the furnace winding. Also, under rather extreme conditions used, the pressure vessel itself would be attacked and damaged by the water vapor.

The charges are placed in the pressure vessel, and the pressure is applied before heating. At the end of the runs the charges are quenched by turning off the power. Cooling is very rapid, and there is little doubt that the equilibrium attained at the high pressures and temperatures is "frozen in" on cooling.

Results

Preliminary results are shown in Figure 1 and Table 1. The change in the temperature of the quartz-tridymite inversion with pressure has been determined and will be reported elsewhere.

The effect of water on liquid silica is most remarkable. The liquidus of cristobalite is lowered by water vapor (Fig. 1) and at pressures above

approximately 400 kg/cm² tridymite melts directly to a hydrous liquid. At water vapor pressures above approximately 1400 kg/cm² tridymite is not formed as a stable phase, and quartz melts example, the liquids of feldspar composition studied by Goranson (1938) would hold about 4.5 per cent water at the same pressures and temperatures.

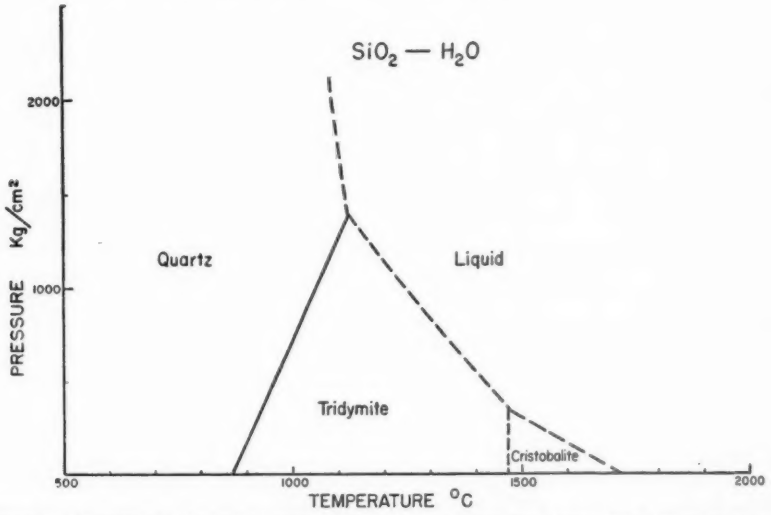


FIGURE 1.—TENTATIVE PRESSURE-TEMPERATURE DIAGRAM FOR THE SILICA-RICH PORTION OF THE SYSTEM SiO_2-H_2O

directly to a hydrous liquid. This represents a considerably greater initial lowering of the liquidus than is produced by water on any of the common rock-forming silicates.

Unfortunately the apparatus failed before we could establish the slope of the *p-t* curve for the quartz $\rightleftharpoons$ liquid boundary with certainty. The slope shown in Figure 1 is probably not greatly in error, but additional experiments are needed to fix the curve exactly.

Charges which were liquid at the temperature and pressure of the run quenched to a clear homogeneous glass. The glass from the quartz (powdered) of run Q2-33D was clear and free of bubbles, suggesting that the hydrous liquid may be considerably less viscous than anhydrous silica glass at 1300°C. (It is unlikely that "dry" silica glass could free itself of bubbles under such conditions.) A water determination gave 2.3 per cent water in the glass, and the dehydration did not destroy its clear homogeneous character. The water content appears to be low for the amount of lowering of the liquidus produced by 1200 kg/cm². For

One of the experiments (Y-2a, Table 1) gave results which appear anomalous. Melting occurred in the capsule which originally contained quartz, but the tridymite and silica glass showed no evidence of a liquid phase. Possibly there was a slight temperature difference between the capsules, but this seems unlikely as they were side by side in the pressure vessel and actually in contact with one another. Another possibility is that the amount of water placed in the capsules was insufficient to fill the interstices of the powdered charges of tridymite and silica glass, and therefore the vapor pressure did not build up to the pressure applied to the capsule. This also appears unlikely. In any event, the fact that the charge which was originally quartz melted indicates that the liquidus temperature for a pressure of 2000 kg/cm² is below 1125°C.

Discussion

Perhaps the most striking feature of the results is the large initial lowering of the liquidus

of silica by moderate pressures of water vapor. The first 1000 kg/cm² water vapor pressure lowers the liquidus nearly 2½ times the amount of lowering produced in albite by a comparable pressure. The next 1000 kg/cm² water vapor pressure apparently produces little additional lowering, presumably because quartz is the primary phase at higher pressures.

It has long been known that water is a potent agent for lowering the liquidus temperatures of certain alkali silicates such as sodium and potassium disilicates (Morey and Fenner, 1917; Tuttle and Friedman, 1948). Also, mixtures of those silicates and silica have a complete solubility relation with water. Knowledge of that solubility relation and of the effect of water on the feldspars, for which the liquidus lowering is less pronounced, has led apparently to the general belief that the alkalies are largely responsible for the tremendous fluxing power of water on silicates. The present study demonstrates that the liquidus of silica itself is profoundly affected by the presence of water vapor under pressure.

If the liquidus of quartz is lowered more rapidly than is shown in Figure 1 with additional water vapor pressure, it will not be so difficult to explain such phenomena as quartz veins and the quartz cores of pegmatites.

The clear bubble-free glass obtained on quenching one run in which powdered quartz was the starting material suggests that the general technique used may be suitable for the preparation of bubble-free silica glass.

Since the solubility of silica in water at the critical end point is extremely small, the critical temperature and pressure of pure water will be only slightly raised by the dissolved material. A second critical end point must therefore lie at extremely high pressures and at a temperature below—perhaps considerably below—1100°C. The pressure, temperature, and composition at the second critical end-point are, of course, unknown but may well lie in a pressure-

TABLE 1.—EXPERIMENTS ON THE SYSTEM SiO_2-H_2O

Experiment Number	Initial Material	Temp. °C	Pressure kg/cm²	Time	Results
Q2-33D	Tridymite	1306	1200	1 hr.	All glass
	Quartz	1306	1200	1 hr.	All glass
	SiO_2 glass	1306	1200	1 hr.	All glass
Q2-33E	Tridymite	1200	1200	1 hr.	All glass
	SiO_2 glass	1200	1200	1 hr.	All glass
Q2-34A	Tridymite	1156	1200	1½ hrs.	Tridymite
	SiO_2 glass	1156	1200	1½ hrs.	Largely tridymite
Q2-34C	Quartz	1000	2000	2 hrs.	Quartz
	SiO_2 glass	1000	2000	2 hrs.	Largely quartz
Q2-34D	Quartz	1075	2000	2 hrs.	Quartz
	SiO_2 glass	1075	2000	2 hrs.	Quartz
*Y-1a	Tridymite	1175	2000	6 hrs.	Largely glass
	Quartz	1175	2000	6 hrs.	Largely glass
	SiO_2 glass	1175	2000	6 hrs.	All glass
*Y-2a	Tridymite	1125	2000	5 hrs.	All quartz
	Quartz	1125	2000	5 hrs.	50% glass ±
	SiO_2 glass	1125	2000	5 hrs.	All quartz

* The writers are indebted to Dr. H. S. Yoder for making these runs in his apparatus. For a description of this equipment *see* Yoder (1950).

temperature range that is accessible with present-day apparatus.

References Cited

Goranson, R. W., 1938, Silicate-water systems: Phase equilibria in the $NaAlSi_3O_8$-H_2O and $KAlSi_3O_8$-H_2O systems at high temperatures and pressures: Am. Jour. Sci., v. XXXV-A, p. 71–91.

—— 1931, The solubility of water in granite magmas: Am. Jour. Sci., v. XXII, p. 481–502.

Kennedy, George C., 1944, The hydrothermal solubility of silica: Econ. Geology, v. 39, p. 25–36.

—— 1950, A portion of the system silica-water: Econ. Geology, v. 45, p. 629–653.

Morey, George W., and Fenner, C. N., 1917, The ternary system H_2O-K_2SiO_3-SiO_2: Jour. Am. Chem. Soc., v. XXXIX, p. 1173–1228.

Morey, George W., and Hesselgesser, James M., 1951, The solubility of some minerals in superheated steam at high pressures: Econ. Geology, v. 46, p. 821–835.

Tuttle, O. F., and Friedman, I. I., 1948, Liquid immiscibility in the system H_2O-Na_2O-SiO_2: Jour. Am. Chem. Soc., v. 70, p. 919–926.

Van Nieuwenburg, C. J., and Van Zon, P. M., 1935, Semi-quantitative measurements of the solubility of quartz in supercritical steam: Rec. Trav. Chim., v. 54, p. 129–32.

Yoder, H. S., 1950, High-low quartz inversion up to 10,000 bars: Am. Geophy. Union. Trans., v. 31, p. 827–835.

College of Mineral Industries, The Pennsylvania State University, State College, Pa.; The Geophysical Laboratory, Washington, D. C.

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 66, PP. 153-154 JANUARY 1955

SELECTED BIBLIOGRAPHY OF STANDARD STRATIGRAPHIC SECTIONS OF NORTH AMERICA

By Fred S. Honkala

A list of standard stratigraphic sections in North America for the various geological periods has been accumulated in advanced historical geology and stratigraphy courses taught at Montana State University. The list is an enlargement of earlier lists by Eardley (1951, p. 2–3) and Krumbien and Sloss (1951, p. 315–316).

The Geological Society of America has published a series of correlation charts which have been prepared under the leadership of the Committee on Stratigraphy of the National Research Council. Originally 12 charts, ranging from Cambrian through Cenozoic, were planned (Dunbar *et al.*, 1942, p. 429). Most of the planned charts have been published, and some periods have been treated in several charts each. The following bibliography lists the charts plus selected references for the geologic time units not covered, including the Precambrian and Quaternary.

Clapp, C. H. and Deiss, C. H., 1931, Correlation of Montana Algonkian formations: Geol. Soc. America Bull., v. 42, p. 673–696, 3 figs.

Grout, F. F., Gruner, J. W., Schwartz, G. M., and Thiel, G. A., 1951, Precambrian stratigraphy of Minnesota; Geol. Soc. America Bull., v. 62, p. 1017–1078, 9 figs.

Howell, B. F., *et al.*, 1944, Correlation of the Cambrian formations of North America: Geol. Soc. America Bull., v. 55, p. 993–1004, 1 pl.

Twenhofel, W. H., *et al.*, 1954, Correlation of the Ordovician formations of North America, Geol. Soc. America Bull., v. 65, p. 247–298, 1 pl., 2 figs.

Swartz, C. K., *et al.*, 1942, Correlation of the Silurian formations of North America: Geol. Soc. America Bull., v. 53, p. 533–538, 1 pl.

Cooper, G. A., *et al.*, 1942, Correlation of the Devonian sedimentary formations of North America: Geol. Soc. America Bull., v. 53, p. 1729–1794, 1 pl., 1 fig.

Weller, J. M., *et al.*, 1948, Correlation of the Mississippian formations of North America: Geol. Soc. America Bull., v. 59, p. 91–196, 2 pls., 7 figs.

Moore, R. C., *et al.*, 1944, Correlation of the Pennsylvanian formations of North America: Geol. Soc. America Bull., v. 55, p. 657–706, 1 pl.

Moore, R. C., and Thompson, M. L., 1949, Main divisions of Pennsylvanian period and system: Am. Assoc. Petroleum Geologists Bull., v. 33, p. 275–301, 2 figs.

Mohr, C. L., 1939, Subsurface cross section of Permian from Texas to Nebraska: Am. Assoc. Petroleum Geologists Bull., v. 23, p. 1694–1711, 3 figs.

Baker, A. A., and Williams, J. Steele, 1940, Permian in parts of Rocky Mountain and Colorado Plateau regions: Am. Assoc. Petroleum Geologists Bull. v. 24, p. 617–635, 7 figs.

King, P. B., 1942, Permian of West Texas and southeastern New Mexico: Am. Assoc. Petroleum Geologists Bull., v. 26, p. 535–763, 34 figs.

Muller, S. W., and Ferguson, H. G., 1936, Triassic and lower Jurassic formations of west-central Nevada: Geol. Soc. America Bull., v. 47, p. 241–252.

Kummel, B., 1953, Regional relationships of Triassic formations in southeastern Idaho and adjacent areas: Intermountain Assoc. Petroleum Geologists, 4th Ann. Field Conf. Guidebook, p. 48–53, 4 figs.

McLearn, F. H., 1953, Correlation of the Triassic formations of Canada: Geol. Soc. America Bull., v. 64, p. 1205–1228, 1 pl., 2 figs.

Imlay, R. W., 1952, Correlation of the Jurassic formations of North America, exclusive of Canada: Geol. Soc. America Bull., v. 63, p. 953–992, 2 pl., 4 figs.

Frebold, H., 1953, Correlation of the Jurassic formations of Canada: Geol. Soc. America Bull., v. 64, p. 1229–1246, 1 pl., 1 fig.

Stephenson, L. W., King, P. B., Monroe, W. H., and Imlay, R. W., 1942, Correlation of the outcropping Cretaceous formations of the Atlantic and Gulf Coastal Plain and Trans-Pecos Texas: Geol. Soc. America Bull. v. 53, p. 435–448, 1 pl.

Imlay, R. W., 1944, Correlation of the Cretaceous formations of the Greater Antilles, Central America and Mexico: Geol. Soc. America Bull., v. 55, p. 1005–1046, 3 pls., 1 fig.

Cobban, W. A., and Reeside, J. B., Jr., 1952, Correlation of the Cretaceous formations of the western interior of the United States: Geol. Soc. America Bull., v. 63, p. 1011–1044, 1 pl., 2 figs.

Imlay, R. W., and Reeside, J. B., Jr., 1954, Correlation of the Cretaceous formations of Greenland and Alaska: Geol. Soc. America Bull., v. 65, p. 223–246, 1 pl., 2 figs.

Lugn, A. L., 1939, Classification of the Tertiary System in Nebraska: Geol. Soc. America Bull., v. 55, p. 1245–1276, 1 pl.

Wood, 2nd, H. E., *et al.*, 1941, Nomenclature and correlation of the North American Tertiary: Geol. Soc. America Bull., v. 52, p. 1–48, 1 pl.

Cooke, C. W., Gardner, J., and Woodring, W. P., 1943, Correlation of the Cenozoic formations of the Atlantic and Gulf Coastal Plain and

Carribean Region: Geol. Soc. America Bull., v. 54, p. 1713–1722, 1 pl.

Weaver, C. E., 1944, Correlation of marine Cenozoic formations of western North America: Geol. Soc. America Bull., v. 55, p. 569–598, 1 pl.

Simpson, G. G., 1947, A continental Tertiary time chart: Jour. Paleontology, v. 21, p. 480–483, 1 fig.

Flint, R. F., 1947, Glacial geology and the Pleistocene epoch: N. Y., John Wiley and Sons, 589 p., 6 pl., 88 figs., (*see particularly* p. 188–304).

Colbert, E. H., *et al.*, 1948, Pleistocene of the Great Plains (a symposium): Geol. Soc. America Bull., v. 59, p. 541–630, 1 pl., 11 figs.

References Cited

Dunbar, C. O., *et al.*, (1942), Correlation charts prepared by the Committee on Stratigraphy of the National Research Council, Bull. Geol. Soc. Am., v. 53, pp. 429–434.

Eardley, A. J., (1951), Structural geology of North America, Harper and Bros., New York, 624 pp., 16 pl., 343 figs.

Krumbein, W. C., and Sloss, L. L., 1951, Stratigraphy and sedimentation: San Francisco, W. H. Freeman and Co., 497 p., 122 figs.

Dept. Geology, Montana State University, Missoula, Montana

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 66, PP. 155-158, 2 FIGS. JANUARY 1955

PALEOECOLOGICAL SIGNIFICANCE OF THE STRONTIUM-CALCIUM RATIO IN FOSSILS AND SEDIMENTS

BY K. TUREKIAN

Introduction

Investigators realize that the ratio of strontium to calcium in carbonate shells and carbonate-bearing sediments is a function of several variables:

(1) the Sr/Ca ratio in the liquid phase from which the solid phase is derived,

(2) the particular polymorph—calcite or aragonite—into which the strontium is incorporated,

(3) the vital effect of the organism,

(4) the temperature,

(5) the salinity of the liquid phase.

Possible interrelationships of these factors have been suggested. However, the importance of salinity as an *independent* variable has not been stressed.

Odum (1951b) suggests that the prime determiner of the Sr/Ca ratio in a particular solid phase (such as a shell) is the Sr/Ca ratio in the water. The relationship between the two is that of a direct proportionality. Fresh water in open lakes and streams has low Sr/Ca, hence the animals secreting shells in this environment will have low Sr/Ca ratios. Marine animals on the other hand secrete shells with high Sr/Ca only because the ocean is higher in this ratio than are fresh waters. The salinity is thus given no independent importance.

Lowenstam (1954) has suggested that the Sr/Ca ratio in organisms with mixed aragonite and calcite tests is a function of the percentages of these two polymorphs. The amount of aragonite in such a test is a function of temperature—the higher the temperature the larger the percentage of aragonite deposited. Since aragonite accepts the large strontium atom more readily than calcite the higher strontium content is expectable in the test deposited in the warmer water. Lowenstam suggests that a variation in salinity may cause a variation in the aragonite/calcite ratio but does not pursue the matter further.

It is the thesis of this paper that an important independent factor in determining the Sr/Ca ratio in shells and sediments may be the salinity of the water.

Florena Shale

Imbrie (1954) has described facies changes in the Florena shale of Kansas. The present paper was initiated by reviewing the data on the variation of the Sr/Ca ratio vertically through a section of the Florena in central Kansas. The Florena is an early Permian calcareous shale (interpreted as shallow-water marine deposition) ranging from 3 to 14 feet thick and bounded on the top and bottom by fusulinid-bearing marine limestones.

The Sr/Ca ratios were determined spectrographically as described by Kulp, Turekian, and Boyd (1952) for two faunal groups and the acid leach of the sediment. The two phyla studied, represented by *Chonetes* shells (Brachiopoda) and crinoid columnals (Echinodermata), deposit exclusively calcite shells and hence eliminate a variation in polymorphic percentages. Very few animal types are represented in the shale that could precipitate any sort of aragonite or composite test. Figure 1 shows the variation of Sr/Ca with height in the section. Where a point is missing the group was absent or very rare in that stratigraphic interval.

The possibility of the variation in the Sr/Ca due to secondary effects, such as ground water, is discounted since it would appear likely that such action would tend to level out differences rather than accentuate them.

Quantification of Odum's Thesis with Respect to Closed or Restricted Basins

Considering the paleogeography of the Permian of Kansas, the first hypothesis to test in order to account for the marked variation in Sr/Ca ratio would be that of deposition in a closed or restricted basin.

Odum has demonstrated by controlled growth of aragonitic gastropod tests that the Sr/Ca ratio of the solid phase is directly related

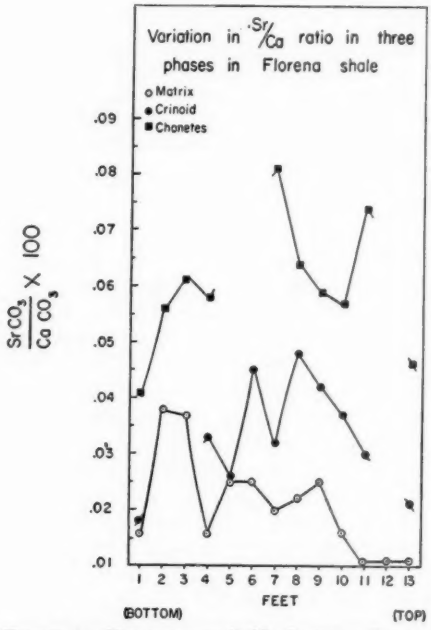


FIGURE 1.—VARIATION IN Sr/Ca RATIO IN THREE PHASES IN FLORENA SHALE, KANSAS

to the ratio in the liquid phase. In a simple statement:

$$R_s = k\,R_l$$

where: R_s and R_l are the Sr/Ca ratios in the solid and liquid phases respectively, and k is a constant characteristic of the system.

In the case of an isolated system where there is continued transfer of calcium from the liquid phase to the solid, there will be an increase of the Sr/Ca ratio in the liquid phase as deposition proceeds if k is less than unity. Consequently the solid phase should also increase progressively reflecting directly the instantaneous ratio of the liquid phase.

This is simply a statement of the classic Doerner-Hoskins logarithmic distribution law for a trace element coprecipitating with a major element. The formal statement of this law is:

$$\frac{dx}{dy} = k\frac{a - x}{a - y}$$

where x and y are the amounts of the two substances X and Y already in the solid phase; a and b are the amounts present in the entire system of X and Y respectively (originally all in the liquid phase). Then the ratio of X to Y in any given increment (expressed as dx/dy) will be related as shown above.

To put this formal expression into working form: Let

A = total number of Sr atoms in a closed basin
B = total number of Ca atoms in a closed basin
q_n = the number of Ca atoms per one Sr atom in the solid phase for the nth increment of Sr removed from the liquid phase.
r = a fixed increment of Sr transferred from the liquid phase to the solid phase.

At the beginning of sedimentation of the following relationship will hold:

$$R_s = \frac{1}{q_0}$$

$$R_l = \frac{A}{B}$$

since all the ions are in the liquid phase. Then according to the Doerner-Hoskins relationship:

$$\frac{1}{q_0} = k\frac{A}{B}$$

For successive increments of strontium transferred (r) from the liquid phase to the solid phase we get:

$$\frac{1}{q_1} = k\frac{A - r}{B - q_0 r}$$

next

$$\frac{1}{q_2} = k\frac{A - 2r}{B - (q_0 + q_1)r}$$

or generally:

$$\frac{1}{q_n} = k\frac{A - nr}{B - r\sum_{j=0}^{n-1} q_j}$$

To evaluate the significance of the above equations let us assume the following highly probable conditions:

(1) In sea water the ratio of Sr/Ca is approximately 10Sr/1000Ca—Odum gives 9.23Sr/1000Ca (1951a)—hence we can substitute $B = 1000A$ in our equation.

(2) A value of k for calcite organisms and sediments may be close to 0.1. (This value is not critical for the argument).

Letting $r/A = \alpha$, since we do not have any idea as to the value of A but can make arbitrary calculations for different values of α (the increment ratio), our equation simplifies to:

$$q_n = \frac{100 - \alpha \sum_{j=0}^{n-1} q_j}{0.1(1 - n\alpha)}$$

We can plot q_n against $n\alpha$, and an approximate linear plot results. In Figure 2 such a plot is made with the increment ratio α chosen as .001. A similar curve would be obtained for other values of α.

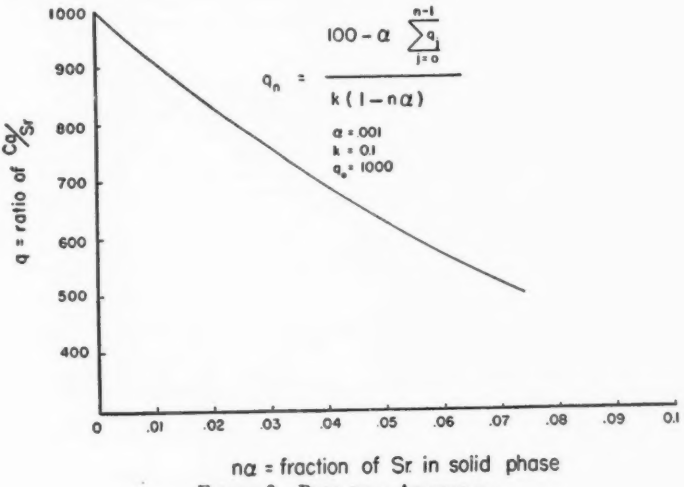


FIGURE 2.—PLOT OF q_n AGAINST $n\alpha$

Application to the Florena Suite

Figure 1 shows that the limits of variation in the Sr/Ca ratio of the various phases studied are:

Phase	Maximum	Minimum	Maximum / Minimum
Chenetes	.081	.041	~2
Crinoid	.048	.018	~2.5
Matrix	.038	.011	~3

In Figure 2 it can be seen that to increase the Sr/Ca ratio in the solid or liquid phase by a factor of two, 7.4 per cent of the total strontium in the system would be found in the solid phase. This corresponds to about 53.8 per cent of the calcium also entering the solid phase.[1] This condition would not allow for the deposition of the normal marine fauna representative of the Florena shale. Consequently Odum's thesis of a simple direct relation between Sr/Ca in the solid to the Sr/Ca in the liquid appears inadequate to explain the observed data.

Effect of Temperature

Both Odum (1950a) and Kulp, Turekian, and Boyd (1952) state that temperature plays only a small role in varying the Sr/Ca ratio in shells. Both papers show data on similar modern faunas from arctic and equatorial marine waters with little or no strontium variation. Lowenstam (1954), however, has demonstrated that there is a Sr/Ca variation that is temperature dependent in organisms having the ability

[1] For the conditions given it follows that the percentage of the total calcium present in the solid phase up to the nth increment of strontium removal is simply: $\alpha \sum_{j=0}^{n-1} q_j$

to vary their aragonite content with temperature.

Such organisms will be the ideal case of variation of Sr/Ca with temperature. His data seem to indicate roughly that in the worm family Serpulidae a variation in strontium content from 0.45 per cent $SrCO_3$ to 1.0 per cent $SrCO_3$ involves a change in aragonite content of the shell relative to calcite from 0 to 90 per cent aragonite. This variation in turn is related to the temperature such that 0 per cent aragonite roughly corresponds to 5° C. or less and 90 per cent aragonite corresponds to 25° C. or more.

All the phases in the Florena shale reviewed in this paper must have been deposited as calcite, hence any variation with temperature might be expected to be subdued. Further, it is extremely improbable that a single locality in central Kansas, continuously occupied by a sea, underwent the extremes of temperature given above.

Conclusion

With the elimination of the above variables it appears that salinity may have played a significant role in determining the Sr/Ca ratio in the Florena suite. Since there is a marked relationship between this chemical indicator and the abundances of various faunal types in the Florena fauna as stated by Imbrie (1954), the possible value of Sr/Ca ratios as an ecological tool to determine changing salinities should not be overlooked.

References Cited

Imbrie, J , 1954, A quantitative lithofacies and biofacies study of the Florena shale (Permian) of Kansas *in* Faunal facies and their lithologic associations (symposium): Soc. Econ Paleontologists and Mineralogists (in press).

Kulp, J. L., Turekian, K., and Boyd, D. W., 1952, Strontium content of limestones and fossils: Geol. Soc. America Bull., v. 63, p. 701–716.

Lowenstam, H. A., 1953, Environmental relations of modification compositions of certain carbonate secreting marine invertebrates: Nat. Acad. Sci. Proc., v. 40, p. 39–48.

Odum, H. T., 1951a, Notes on the strontium content of sea water; Science, 114, p. 211–213.

—— 1951b, The stability of the world strontium cycle: Science, v. 114, p. 407–411.

Lamont Geological Observatory, Palisades, New York

Lamont Geological Observatory Contribution No. 121